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PLEISTOCENE RESEARCH (A REVIEW BY THE MEMBERS OF THE COMMITTEE ON INTERRELATIONS OF PLEISTOCENE RESEARCH, NATIONAL RESEARCH COUNCIL). CONTAINS 12 PAPERS. R. F. Flint, *Chairman*

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DEVELOPMENT OF LAKE SUPERIOR SOFT IRON ORES FROM METAMORPHOSED IRON FORMATION

By Stanley A. Tyler

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Abstract

Local metamorphism, probably of the hydrothermal type, adjacent to dikes, sills, quartz-adularia pegmatite veins, quartz-tourmaline veins, and quartz veins has induced secondary iron silicates such as grunerite, minnesotaite, and stilpnomelane to form at the expense of the primary minerals, chert, siderite, and greenalite. The elimination of chert through the development of iron silicates is favored by an original high iron content of the rock, by primary textures which allow the reacting minerals to be in close proximity, by the formation of secondary low iron and high silica minerals, and by the introduction of iron.

Oxidation of either the original siderite or greenalite iron formation or a facies composed of secondary iron silicates and chert produces the ordinary red or yellow cherty iron formation. The soft ores, both limonite and hematite, result from the oxidation and leaching of that facies in which the chert has been largely or completely combined with iron to form iron silicates. Chert has not been removed as such in the quantities hitherto considered necessary for the formation of ore.

Acknowledgments

The author has studied the soft ores of the Lake Superior region for the past 11 years in connection with consulting work carried on for the Jones and Laughlin Steel Corporation. Nearly a thousand thin sections of the various phases of the iron formation and ore were examined during the course of the study. Chemical analyses made by W. R. Pasich were furnished by the Company. Acknowledgments are gratefully made to Mr. L. P. Barrett, chief geologist for the Company, who realized that a better understanding of the origin of the ores would lead to more successful prospecting. Appreciation is also expressed to C. K. Leith, J. W. Gruner, L. P. Barrett, C. E. Dutton, L. M. Scofield, and R. W. Marsden for constructive criticism of the manuscript.

Introduction

The origin of the Lake Superior soft ores has received the attention of many students since the problem was first discussed by Foster and Whitney (1851, p. 69). Two general but diverse theories have emerged from the wealth of factual information accumulated over the years by Federal, State, and private surveys. Van Hise and Leith (1911, pp. 499–570) supported the theory that the iron ores owe their secondary concentration to oxidizing and leaching solutions moving downward and laterally from the erosion surface. Gruner (1926a, p. 629–644) advanced the concept that the oxidizing and leaching solutions were hydrothermal in origin. He later modified this by suggesting that the solutions were largely meteoric in origin, but that emanations rising from a magmatic source furnished the heat and produced the upward circulation (1937, p. 121–130).

The advocates of both theories seem to be in general agreement on the following points:

1. The iron formations were originally composed dominantly of chert, with important but subordinate quantities of siderite and greenalite, and perhaps some ferric oxide.

2. Oxygen-bearing solutions have oxidized the ferrous minerals, leached the silica from the ferrous silicates, and removed the chert, forming a residual product, the ore.

Leith and Gruner agree that the removal of large quantities of chert by solution was necessary for ore formation. In contrast, the author advances the concept that the solution of chert has been of less importance and that the soft, porous ores were formed by oxidizing and leaching solutions acting upod a metamorphic facies composed dominantly of secondary iron silicates, with little or no chert. This point of view necessitates a re-examination of the character and distribution of the metamorphic facies of the iron formations, as well as the nature of the ferruginous cherts and the ore.

Metamorphic Facies of the Iron Formations

General statement

According to Van Hise and Leith (1911, p. 500), "... the original constituents of the iron-bearing formations were dominantly cherty iron carbonate and iron silicate (greenalite), with minor amounts of hematite and magnetite...." These constituents were either deposited as original precipitates in the Precambrian seas or produced by diagenesis. The cherty iron carbonate or greenalite facies are now but remnants of the original iron forma-

tions which have escaped subsequent metamorphism. Deep burial with associated high temperatures and pressures has favored regional metamorphism throughout the Lake Superior region. Contact metamorphic zones located adjacent to the larger intrusives, such as the Duluth gabbro on the eastern Mesabi, Mellen gabbro on the western Gogebic, and Republic granite on the western Marquette have been described by many workers. Some of the more recent discussions of these areas are those of Grout and Broderick (1919, p. 8–10), Aldrich (1929, p. 95–97), and Lamey (1934, p. 248–263). The zones of local metamorphism which occur adjacent to dikes, sills, quartz-adularia pegmatite veins, quartz-tourmaline veins, and quartz veins have not been emphasized or adequately described.

Minerals and Their Origin

General statement.—The metamorphic facies of the iron formations where unaffected by oxidation is a grayish-green rock usually consisting of one or more iron silicates, magnetite, and sugary quartz. Grunerite, stilpnomelane, or minnesotaite are usually abundant; greenalite, siderite, and chert occur in variable but subordinate quantities.

Grunerite.—Grunerite is most extensively developed on the eastern Mesabi, the western Gogebic, and the western portion of the Marquette district, where it is associated with other amphiboles, magnetite, chlorite, garnet, fayalite, and pyroxene. The mineral association, the rather coarse texture of the rock, and the proximity to large intrusives indicates that grunerite occurs in a rather high-grade metamorphic facies.

Stilpnomelane.—This mineral was first recognized in the Lake Superior district by Grout and Thiel (1924, p. 228–231) in veins cutting the iron formation on the Mesabi and Cuyuna ranges and in slates on the Cuyuna range. Gruner (1944a, p. 291–298) reports the mineral to be very common in the Biwabik formation on the Mesabi range. Ayers (1940, p. 432–437) was the first to note the presence of stilpnomelane in the ferruginous upper Huronian slates of the Crystal Falls district, and Pettijohn (1946) reports that it is abundant in the hanging-wall slates of that district. Stilpnomelane is very abundant in the metamorphic facies of the iron formation in the eastern part of the Gogebic and in the Marquette district.

Gruner (1946, p. 12) postulates that stilpnomelane may have formed from a colloidal gel, with minnesotaite or greenalite or quartz and siderite developing from the remaining constituents of the gel. According to this concept, minnesotaite and stilpnomelane would be considered as primary constituents with greenalite. Hallimond (1924, p. 193–197) describes the occurrence of stilpnomelane in a metamorphic facies of iron formation, adjacent to greenstone dikes in North Wales, and Ayers (1940, p. 432) states that in the Crystal Falls district, "It was formed by hydrothermal reaction accompanying the emplacement of quartz-adularia pegmatite veins." Turner and Hutton (1935, p. 1–8) discussing the origin of stilpnomelane in graywacke schists from New Zealand, state: "Thus in the schists of western Otago stilpnomelane has crystallized as a stable product of low grade metamorphism, in which shearing stress, and low temperature, aided to an unknown extent by the action of dilute solutions has been the dominant controlling factor."

Stilpnomelane is extensively developed in iron formation and slates, adjacent to basic dikes and sills, quartz-adularia pegmatite veins, quartz-tourmaline veins, and quartz veins on the eastern Gogebic, the Marquette range, and in the Crystal Falls district. It occurs in a much lower-grade metamorphic facies than grunerite, but has often been mistaken for grunerite. Many areas which have been classified as grunerite-magnetite schists are actually stilpnomelane-magnetite rocks.

Minnesotaite.—Another iron silicate, minnesotaite, was recently discovered by Gruner (1944b, p. 363–372) on the Mesabi and Cuyuna ranges. Minnesotaite was identified by the author from the Ironwood and Negaunee formations on the eastern Gogebic and Marquette ranges, where it occurs in association with stilpnomelane and magnetite. This association indicates that the mineral probably forms in a low-grade metamorphic facies under conditions similar to stilpnomelane.

Magnetite.—This mineral is abundant in some metamorphic facies of the iron formations, but is absent in others. Magnetite seems to

have developed from siderite or greenalite, when oxygen was available from any source during metamorphism. If sufficient oxygen were present to change all the siderite and greenalite to magnetite, the final product would consist of magnetite and chert. Allen and Barrett (1915, p. 72) describe this metamorphic type from the eastern portion of the Marenisco range. Where oxygen was deficient, the iron combined with the chert to form iron silicates exclusively. Much of the Goose Lake iron formation and parts of the Negaunee iron formation in the Marquette district are composed largely of secondary iron silicates, but contain little if any magnetite. The more normal situation seems to have been the development of some magnetite from the siderite and greenalite, with the remainder of the iron combining with chert to form iron silicates.

Chert (quartz).—Chert, or finely crystalline quartz, is an abundant primary constituent of iron formation. It tends to be eliminated in the metamorphic facies through the formation of iron silicates, but what does not combine with iron serves as an indicator of the grade of metamorphism reached. Aldrich (1929, p. 90–91) noted this relationship on the western Gogebic: "The particular iron mineral present is somewhat closely correlated with the texture of the quartz. Thus, the uniformly coarsest texture is always associated with the silicates, and usually has magnetite and hard blue hematite in association. The most dense-textured quartz (chert) is associated only with carbonate." Although the chert is very fine-grained in the original iron formation, the grain size increases rapidly with the grade of metamorphism; at the grunerite stage it often assumes the size and granularity of sugar and is commonly referred to as "sugary quartz." The grain size reaches a maximum of about a fourth of an inch close to some of the larger basic intrusives, such as those at the Lothian and Iron Mountain Lake mines in the Marquette district.

Factors That Influence the Elimination of Chert through the Formation of Iron Silicates

General statement.—The location, shape, and size of ore bodies is determined, according to the author's concept of soft ore formation, by the oxidation and leaching of those portions of the metamorphic facies of the iron formations in which the chert has been largely or completely combined with iron to form secondary iron silicates. These iron silicates have formed in response to both regional and local metamorphic environments. Under the regional environment, the extent to which the chert is combined with the iron depends in a large measure upon the effective iron content and the texture of the original iron formation, and the composition of the secondary silicates formed.

Local metamorphism, which is superimposed upon and therefore intensifies the regional effect, is most pronounced adjacent to dikes, sills, quartz-adularia pegmatite veins, quartz-tourmaline veins, and quartz veins. The factors which favor the formation of iron silicates under the regional environment are important here, but the elimination of the chert is usually more complete. The local source of heat, presence of hydrothermal solutions, and the migration or introduction of iron also seem to have promoted the reaction of the chert with the iron.

Effective iron content.—The development of magnetite from siderite or greenalite, before or when the iron silicates form, isolates a portion of the iron in a mineral which appears to be more or less stable in the presence of chert, at least in the lower grades of metamorphism. This decreases the iron available for combination with the silica in the chert and may lead to incomplete elimination of the chert. The same result is also produced, without the formation of magnetite, when the iron content of the rock (either original or introduced) is less than that necessary to satisfy most of the silica in the chert.

Primary textures.—The texture of the original iron formations seems to have had an important bearing upon the completeness of the mineral reactions. Reactions of the silica of chert with iron seems to have been more complete in beds where siderite was disseminated throughout the chert than where beds of chert and siderite alternated. Likewise thin alternating beds of chert and siderite lead to more complete reaction than thick alternating beds.

The granular texture which is so well developed in the iron formation on the Mesabi range and at certain horizons on the Gogebic range favors the formation of iron silicates,

because of the intimate physical relationship of siderite, greenalite, and chert. Any one of these minerals may occur either as granules or as the matrix to granules of one of the other minerals.

The profound influence of texture and original iron content upon the development of iron silicates at the expense of chert probably explains why the chert at certain stratigraphic horizons in the iron formation is almost completely combined with the iron, whereas adjoining beds contain chert, siderite, greenalite, and secondary iron silicates.

Secondary iron silicates.—The specific iron silicate or silicates that develop determine to a large extent whether or not the elimination of the chert will be complete. The formation of secondary silicates with a low iron and high silica content are more effective in removing the chert than high-iron and low-silica minerals. Iron-rich silicates formed at low temperatures may take on more silica from the chert during metamorphism to form secondary silicates with a lower iron content. Thus greenalite which contains about 39 per cent iron may be observed in thin sections in all stages of alteration to minnesotaite. Chert is eliminated in this process.

We may consider a facies of the original iron formation which consisted essentially of siderite and chert. If stilpnomelane—21 to 29 per cent iron—or minnesotaite—27.5 per cent iron—forms, the chert may be changed completely to iron silicates with an original iron content as low as 20 to 23 per cent. Chert in excess of that required to form iron silicates recrystallizes to form sugary quartz.

If grunerite of the composition

$$Fe_7(Si_4O_{11})_2(OH)_2$$

—39.6 per cent iron should form from a siderite chert rock, complete combination of the chert with iron would require an original iron content slightly in excess of 30 per cent. Thus elimination of the chert by an iron-rich silicate of this type would be confined to the richer phase of siderite iron formation or to areas which had received contributions of iron from elsewhere.

Sills and dikes.—Basic sills and dikes occur within the iron formation on the Marquette and Gogebic ranges. On the Marquette range, numerous thin (1–50 feet) concordant bodies of basic igneous rock may be tuff beds, lava flows, or sills. The author agrees with Derby (1929, p. 731–733) that the thick, tabular bodies in the Negaunee formation southwest of Negaunee and south of Ishpeming are sills; the pronounced metamorphism both above and below these basic sheets clearly indicates their intrusive nature. A recent drill hole which penetrated a 300-foot sill located in the NW¼ of section 7, 47–26 shows that stilpnomelane, minnesotaite, and magnetite are developed in the Negaunee formation for 300 feet below the sill. The Negaunee formation immediately above the sill has a greater quantity of iron silicates and less chert than that below, but oxidation and erosion prevent an accurate estimate of the upward limit of the development of iron silicates. Numerous drill holes which penetrated the top of the sill show that the amount of iron silicates formed at the expense of the chert and the coarseness of the silicates varies at points equidistant from the sill. In some areas, a zone of iron silicates with subordinate amounts of chert borders the sill; at other points the silicate areas are irregular and sporadic. Iron formation consisting of chert with subordinate amounts of silicates and siderite may lie between the sill and areas composed largely of iron silicates. Tabular blocks of the Negaunee formation included within the sill may consist entirely of iron silicates, magnetite, siderite, and chert, or of chert and siderite. Metamorphic facies of the iron formation associated with sills may be observed at many localities in the Marquette district. The Negaunee formation, located above and included in the top of the thick sill at the Lothian mine south of Ishpeming, and the Goose Lake iron formation, which lies below a thick sill in the eastern part of the district, are particularly well exposed.

A diabase sill 1000 feet thick occurs about 1200 feet above the base of the Ironwood formation on the eastern Gogebic. This sill extends from near the center of section 15, 47-44 to section 11, 47-45, where it passes unconformably beneath the Keweenawan lava flows. The chert of the Ironwood formation has been partially altered to minnesotaite and stilpnomelane, for 400–500 feet below this sill. Recent drilling in section 8, 47-44 shows that a slaty facies of an iron formation above the sill is composed largely of iron silicates for a stratigraphic distance of 1000 feet. The un-

conformity at the base of the Keweenawan eliminates the upper part of the Ironwood at this point.

Intense development of iron silicates is also associated with the basic dikes on the Marquette and Gogebic ranges. Certain dikes appear to have been more effective than others in promoting the formation of iron silicates. In section 12, 47-45 on the eastern Gogebic, the formation of minnesotaite and stilpnomelane from sideritic chert extends for only 1 or 2 feet on either side of a 20-foot dike; the iron formation adjacent to a near-by dike of approximately the same thickness has been altered almost entirely to iron silicates for 20 feet above the dike, but shows only a slight development of silicates below. The location of areas composed largely of secondary iron silicates is almost invariably related to the attitude of the dike. Secondary iron silicates are usually more abundant on the upper side of inclined dikes, as in the example just cited. Iron formations adjacent to vertical dikes, however, usually have the iron silicates developed to an equal degree on either side of the dike. An example of this relationship may be observed near the north side of the New Volunteer open pit, at Palmer, Michigan, where iron silicates are developed in the Negaunee formation for 20-30 feet, on either side of a vertical dike.

The iron formation adjacent to thin dikes often has a larger proportion of iron silicates than that near thin sills. The dikes apparently were feeding channels, and their supply of heat and hydrothermal solutions was constantly replenished, while that of the sills was dissipated. Stutzer (1940, p. 299) quotes Uthemann to the effect that dikes produce a more profound and widespread effect upon the coals of Lower Hesse than sills.

Heat effects alone should extend for equal distances above and below a flatly-dipping, or on either side of a steeply-dipping intrusive if the heat conductivity of the country rock was more or less uniform. However, elimination of the chert through formation of iron silicates seems to be more complete above intrusives in iron formations rather than below. This suggests that hydrothermal solutions played an important role. These solutions probably were derived in part from the intrusive and in part from the deeper-lying magmatic source. McFarlane (1929, p. 2–5) in describing the effects of basic intrusives upon coal beds in Colorado states that: "In the central area of a sill anthracite was found as far below the sill as above it, showing that heat penetrated downward as readily as upward. However, the zone of discernible metamorphism extends upward considerably farther than downward. . . ."

Taliaferro (1943, p. 109–219) states that hydrothermal solutions were an important factor in the formation of the glaucophane and actinolite schists associated with ultra-basic intrusives in the Knoxville sandstone, shales, cherts, and basalts of California. His descriptions of the alteration of the Knoxville are particularly pertinent, for the mode of occurrence of the metamorphic facies seems to be almost identical to that found in the iron formations of the Marquette and Gogebic ranges. He states (1943, p. 162)

> "Most of the schists are developed from sediments or volcanics forming the roof of the sill or from included leaves within the sill. If any schists are formed by the intrusion, the greatest development occurs in the roof, which is to be expected since the easiest avenue of escape of the volatiles and hot solutions would be upward. Small patches of schists also occur on the bottom of sills."

He also notes (1943, p. 165) that "There is great irregularity in the distribution of schists even above the sills. Some of the schist belts are continuous and coincident with the sill but more commonly the distribution is irregular and there are small areas of schist completely surrounded by unaltered sediments even on the upper side of the sill."

Quartz-adularia pegmatite veins.—The association of secondary iron silicates with quartz-adularia pegmatite veins was first noted by Ayers (1940, p. 432–435) in the Crystal Falls district of Michigan, where stilpnomelane is developed in the ferruginous Upper Huronian slates in great abundance. The veins at this locality are numerous and irregular, making it difficult to relate specific areas of intense silication to individual veins. The coarsest stilpnomelane occurs in and immediately adjacent to the veins; the slates for tens of feet surrounding the veins are composed almost entirely of this mineral.

Quartz-adularia pegmatite veins also occur

near the base of the Negaunee iron formation at the Rolling Mill mine in the Marquette district and at a number of localities in the upper part of the Ironwood formation on the eastern Gogebic. These occurrences all show intense development of iron silicates over a wide area adjacent to the vein, with the coarsest silicates developed close to or in the veins. This development may be due to the metamorphic effect of hydrothermal solutions which rose along the vein channelways and permeated the iron-bearing formation.

Quartz-tourmaline and quartz veins.—The genetic relationship of the quartz-adularia pegmatite, the quartz-tourmaline, and the quartz veins is not known, but they may belong to the same sequence. Iron silicates are often developed in great abundance adjacent to quartz veins bearing tourmaline and in some cases in the vicinity of quartz veins, whereas other quartz veins have apparently had little effect upon iron formation. Perhaps the temperature or the nature of the hydrothermal solutions traveling along these channelways differed materially. Iron silicate zones 20–30 feet wide bordering some of the veins indicate that solutions have had a potent effect upon iron formation and that few channelways were needed to alter large areas.

Introduction and migration of iron.—The question arises whether iron has migrated within or has been introduced during metamorphism. Either of these two processes might lead to the elimination of the chert through the formation of iron silicates in rocks which contained little or no original iron. The introduction of iron from a source outside the iron formations is difficult to prove. The migration of iron is evident, but the source of this iron and the distance of migration often are in doubt.

The formation of iron silicates from thin alternating beds of siderite and chert requires that the iron migrate for only a fraction of an inch, but the formation of silicates from thick-bedded chert and siderite may involve migration of the iron over a distance of several feet. The thin-bedded iron formation which crops out immediately east of the main workings at the Kloman mine near Republic, Michigan provides an excellent opportunity to study the results of this process. Here remnants of iron formation composed of chert and iron silicates are surrounded by iron formation consisting dominantly of iron silicates. The sharp borders of the remnants outline the sharply defined front of the wave of silication. Beds of almost pure chert within the remnants are altered to a silicate rock beyond the border.

Evidence for the migration of iron may be observed in the thick-bedded cherts of the Negaunee and Bijiki iron formations in the western part of the Marquette district. At the Michigamme mine, veinlets of grunerite, which appear to be following joints, extend into and ramify through almost iron-free, massive cherts. Between the Marine mine and the town of Champion, the lean, white cherts of the Bijiki iron formation contain veins and irregular areas of iron silicates cutting across the bedding of the massive cherts. The extensive formation of iron silicates in the extremely lean chert beds indicates that the migration of iron, probably in the ferrous state, must have been an important factor in the elimination of the chert.

If hydrothermal solutions entered the iron formations during metamorphism and aided in the transfer of iron, one might expect the solutions to have come into contact with the basic dikes and sills which occur on the Marquette and Gogebic ranges. The fresh and unaltered intrusives are fine to coarse-grained diabases consisting dominantly of augite, or hornblende after augite, and lath-shaped plagioclase feldspar with minor amounts of ilmenite, magnetite, and in some cases olivine. Irving and Van Hise (1892, p. 271–274, 355–359) and Van Hise and Bayley (1897, p. 487–524) have described both the fresh and altered phases of these rocks. The basic intrusives have been altered rather extensively to chlorite schists and massive chlorite rocks. The alteration is characterized by the replacement of hornblende, augite, and feldspars by chlorite or other ferrous silicates, with the development of minor amounts of epidote, calcite, and sericite from a portion of the feldspar. This alteration appears in thin section in all phases of development from an incipient stage, where the chlorite comes in along cleavage, zones or cracks in the feldspar, to complete replacement of the feldspar. Stilpnomelane appears in a more advanced phase of the alteration as small pleochroic brown to greenish flakes and needles replacing the chlorite. The stilpnomelane may become more

abundant than the chlorite in an advanced phase of the alteration. Turner and Hutton (1935, p. 7) describe stilpnomelane replacing chlorite in schists from New Zealand.

The development of chlorite and stilpnomelane in dikes and sills is accompanied by an increase in total iron and a decrease of silica in relation to iron and alumina. The iron content of the unaltered diabase ranges from 8 to 13 per cent; the chlorite phase ranges from 19 to 30 per cent iron, without visible magnetite in thin section. If magnetite is present, the iron content may be even higher. The surprisingly high iron content of the chloritized diabase approximates that of average iron formation and explains why dikes commonly cannot be distinguished from iron formation on the basis of the iron analysis.

Thin dikes and sills are often chloritized throughout, whereas the thicker bodies may be altered only near the border. The diabase texture is usually obscure or entirely lacking in the sheared facies, the rock becoming a chlorite schist. The diabase texture may still be preserved in the more massive facies, although the original minerals are entirely replaced by iron silicates. Some of the schists, particularly those in which stilpnomelane is abundant, resemble the metamorphic facies of iron formation very closely, but an analysis for titanium or the examination of a thin section for ilmenite or leucoxene will usually serve to distinguish them. The basic dikes and sills contain in excess of 1 per cent titanium dioxide; iron formation contains less than two-tenths of 1 per cent.

It seems evident that iron has been contributed to the basic intrusives after emplacement by some form of hydrothermal activity, either iron-bearing meteoric waters heated by the intrusives or hydrothermal emanations rising from greater depths.

Relation of Iron Silicates to Stratigraphic Horizons

Stratigraphic horizons with a high iron content, a granular texture, thin alternating beds of chert and siderite, or zones located adjacent to sills may be altered almost completely to iron silicate rock. This may be termed stratigraphic alteration. Sills, dikes, quartz-adularia pegmatites, quartz-tourmaline veins, and the introduction or migration of iron may aid stratigraphic silication in some cases, and in others may promote the development of iron silicates with little regard to the bedding. This may be termed nonstratigraphic alteration. At a given locality, the development of iron silicates may be stratigraphic, nonstratigraphic or both.

General Effects of the Development of Silicates upon Iron Formation

General statement.—The development of iron silicates at the expense of chert has in general rather profoundly effected the iron formations, to the extent that the rocks were sometimes not recognized as iron formation. Principle changes besides the development of secondary minerals have been an increase in grain size, an increase in density, development of porosity in some cases, and a tendency for the rock to become more massive or slaty in appearance.

Grain size.—No relation has been observed between the abundance of minnesotaite, stilpnomelane, or grunerite and grain size. The iron silicates range from microcrystalline in the vicinity of Negaunee, Michigan to coarsely crystalline, with individual grains over an inch long, in the western portion of the Marquette district. A marked tendency for grunerite to become coarser-grained than stilpnomelane or minnesotaite may be because grunerite develops in a higher-grade metamorphic facies, where there is more freedom of growth.

Porosity and density.—The porosity of metamorphic iron formation is usually considered to be very low; however, studies of drill core indicate that some portions are decidedly porous and permeable. The combination of siderite and chert to form stilpnomelane or minnesotaite involves a volume decrease. The exact volume change depends upon the amount of carbon dioxide eliminated through the formation of iron silicates, and this in turn depends upon the proportion of the different silicates developed. Recrystallization of the rock under high confining pressures probably tends to reduce the volume as a whole, thereby increasing the density and decreasing the porosity. Lower pressures would favor the retention of porosity.

Bedding.—The development of iron silicates

in both the carbonate and chert beds tends to obscure the pronounced banding of siderite-chert iron formation. Metamorphic iron formation usually ranges from massive to slaty in appearance depending upon whether or not the banding has been partially or largely destroyed by recrystallization. Thin-bedded metamorphic iron formation on the eastern Gogebic resembles a true slate so closely that many workers have failed to distinguish these two rocks properly. This error usually arises in areas where metamorphic iron formation contains little magnetite.

Origin of the Ferruginous Cherts and Slates

General Statement

Iron oxide is a prominent to dominant constituent of the ferruginous cherts, ferruginous slates, and the ore. The ferruginous cherts consist of the more massively bedded varieties of iron formation, with thick individual chert beds alternating with beds of iron oxide. The chert beds usually contain some iron oxide and the iron oxide beds some chert. The iron oxide is usually hematite, but limonite, martite, or magnetite may be present. The term ferruginous slate, as used in the Lake Superior region, refers to thinly laminated alternating beds of iron oxide and chert, but in some cases is applied to thinly-bedded iron formation of any type. Argillaceous beds with a high alumina content, are rare, except at certain horizons within the iron formations.

Ferruginous Cherts and Slates Derived from Siderite-greenalite-chert Facies

It has been generally assumed that ferruginous cherts and slates are the oxidized equivalents of the siderite, or greenalite—chert facies of the iron formations. This conclusion is based upon specific localities where siderite or greenalite and chert rocks grade laterally into ferruginous cherts or slates. Rhombohedral, negative pseudomorphs of siderite, either empty or partially filled with limonite or hematite, which occur in fine-grained ferruginous cherts suggest that a portion of the ferruginous cherts and slates was derived from original sideritic iron formation. The quantitative importance of this source is not certain.

Ferruginous Cherts and slates Derived from the Metamorphic Iron Formation

General statement.—Oxidation of metamorphic iron formation also produces ferruginous cherts and slates which, superficially at least, resemble those derived from the original iron formation. These ferruginous cherts and slates pass imperceptibly into the metamorphic facies of the iron formations; the change may be observed in outcrop, hand specimen, and thin section. The zone of partial oxidation is rather narrow, forming a peripheral band adjacent to areas of metamorphic iron formation. Van Hise and Bayley (1897, p. 369) recognized that metamorphic iron formation graded into the oxidized phase: "Where weathering has been active, the ferruginous slates and cherts are found rather than grunerite-magnetite-schists; the grunerite-magnetite-schists where weathered have been partly transformed into the ferruginous slates and cherts. . . ." Van Hise and Bayley probably did not realize the extent of this alteration for they assume that most of the ferruginous cherts and slates were derived from the original greenalite or sideritic cherts.

The normal gradation of metamorphic iron formation into the oxidized facies was not generally recognized because most workers failed to distinguish between fine-grained metamorphic iron formation and the original cherty iron carbonate. The iron formation exposed near Goose Lake, in the eastern portion of the Marquette district, has often been classified in the field as cherty iron carbonate. Thin sections show that siderite is very rare and that the material is dominantly a nonmagnetic, metamorphic facies of the iron formation. Van Hise and Leith (1911, p. 263) describing a second area in the Marquette district state: "The sideritic slates are most abundant in the valleys between the greenstone masses in the large area south of Ishpeming and Negaunee." Van Hise and Bayley (1897, p. 336–337) cite a chemical analysis from this area to illustrate the abundance of siderite, but they record the presence of actinolite and grunerite in addition to siderite and chert. The reader is left with the impression that this area is composed largely of

sideritic slates. The rocks are predominantly silicate iron formation, mostly of the magnetic variety, whereas iron formation consisting of siderite and chert with subordinate iron silicates is fairly common only in the southern part of the area near the New Volunteer mine, and the siderite-chert facies is rare. The difficulty experienced by many in distinguishing fine-grained metamorphic iron formation from sideritic cherts and slates leads one to suspect that these two rocks have often been confused in lithologic descriptions and field mapping.

Iron silicates.—Spurr (1894, p. 259) and Leith (1903, p. 278) agree that the primary iron silicate greenalite (identified as glauconite by Spurr and later found to be a new mineral by Leith) has altered through loss of silica to limonite and hematite. Likewise Gruner (1946, p. 97) has shown that the secondary silicates stilpnomelane and minnesotatite alter to goethite and hematite. Grunerite from the Bijiki formation in the western part of the Marquette district may be seen in all stages of alteration to goethite and hematite.

Gruner (1946, p. 97) states, "When the silicates, minnesotaite and stilpnomelane, break down they form at first a dirty green to yellowish soft earthy material, which may have the shape of the original needles. A number of X-ray photographs of these needles give a few poor lines which coincide with equally poor lines of montmorillonite or nontronite. As nontronite is a very poorly crystallized ferric silicate with much H_2O, and a layer structure similar to minnesotaite and stilpnomelane such a transition from fresh silicates to goethite by way of nontronite is not at all surprising." Ayers (1940, p. 433–434) also reports the presence of nontronite in veins in the oxidized iron formation on the Marquette, Menominee, and Gogebic ranges. The wide distribution of this mineral in the iron formation suggests that the oxidation and leaching of iron silicates has been extensive.

On the eastern Gogebic and Marquette ranges, the transitional zone from metamorphic to limonitic iron formation is usually marked by a yellowish-green phase. Nontronite was tentatively identified by means of X-ray powder photographs of this material, but the lines are so poorly developed that positive identification was impossible. Gruner (1946, p. 99) suggests that goethite is the normal residual product of the iron silicates on the Mesabi, but the author's observations on the Gogebic and Marquette ranges are that the iron silicates alter to soft red hematite on a far more extensive scale than to goethite.

Kaolinite.—Gruner (1946, p. 22, 98–99) has identified kaolinite associated with limonite and hematite as a residual product formed by the oxidation and leaching of stilpnomelane on the Mesabi range. The persistent occurrence of kaolinite in the ferruginous cherts and slates from the Marquette and Gogebic ranges may be of this origin.

Martite.—The variety of hematite known as martite is a normal constituent of the ferruginous cherts and slates of the Lake Superior region. Although the presence of this mineral has often been recorded, its abundance has not been fully appreciated. Quantitatively it sometimes exceeds limonite and the earthy and specular forms of hematite. The term martite is applied to hematite in octahedral or dodecahedral crystals pseudomorphous after magnetite. Sosman and Hostedter (1918, p. 409–433) suggest that a continuous series exists from magnetite to isometric Fe_2O_3, but Gruner (1926b, p. 393) states that, "No solid solutions of magnetite and hematite seem to exist in nature." The martite occurring in the Lake Superior iron formations seems to have been derived entirely from magnetite, for all gradations from magnetite to martite with magnetite cores, to octahedral hematite (martite) may be observed in polished and thin sections. Wienert (1933, p. 68–74) describes an example of this type of alteration from the Marquette district and concludes that the martite is an oxidation product of magnetite. The prevalence of martite in the ferruginous cherts and slates indicates that magnetite was an abundant mineral prior to the oxidation of these rocks. Magnetite is a common mineral in some of the metamorphic facies of the iron formations today, and there seems to be little reason to believe that it ever was an abundant primary mineral in the siderite and greenalite cherts. Grout and Broderick (1919, p. 45–46) have expressed the same opinion regarding the magnetite on the Mesabi range. Thus it seems reasonable to conclude that the abundance of martite in the ferruginous cherts and slates in-

dicates that these rocks are the oxidized equivalents of a metamorphic facies of the iron formation.

Texture.—The texture of the chert in the oxidized iron formation provides additional evidence that these rocks are often the oxidized equivalents of metamorphic iron formation. Chert in the original siderite or greenalite iron formation, and the oxidized facies derived therefrom, is fine-grained, but the chert associated with martite, especially when adjacent to dikes, sills, quartz-adularia pegmatite veins, quartz-tourmaline veins, quartz veins, or in areas peripheral to metamorphic iron formation is likely to be coarse-grained. Quartz grains ranging in size up to a quarter of an inch are common in oxidized iron formation adjacent to some basic intrusives. Coarse quartz of this type may be seen at the Ogden siliceous pit, and at the Mitchell Mine south of Ishpeming, Michigan. The chert in some metamorphic facies of the iron formation has acquired through recrystallization a texture and grain size which reminds one of sugar. This texture is well developed in both the metamorphic and the oxidized phase of the iron formation on the western Marquette and eastern Marenisco ranges. The sugary-textured quartz is sometimes very friable and closely resembles the wash ores of the western Mesabi.

The Bijiki iron formation because of its coarse texture offers an excellent opportunity to study the transition from the metamorphic to the oxidized facies of the iron formation. Where unoxidized the Bijiki consists of coarse sheaf-like or radial growths of grunerite associated with magnetite and sugary quartz. The original texture of the silicates is perfectly preserved even to the most minute detail in the oxidized, limonite or hematite facies. All stages in the transition from grunerite to goethite or hematite may be observed.

Summary.—The mineral, textural, and field evidence indicates that ferruginous cherts and slates were derived on a large scale from a metamorphic facies of the iron formation, consisting of chert and secondary iron silicates. Oxidation of the original iron formation probably was a subordinate source. The oxidation which has so profoundly changed the appearance of these rocks is discussed with the formation of the ores.

Occurrence and Origin of the Soft Ores

General Statement

The concept that the soft ores were derived from metamorphic iron formation by solutions, which oxidized the ferrous silicates and leached the silica, is opposed to the generally accepted view that metamorphic iron formation is distinctly unfavorable to ore formation. Van Hise and Leith supported the latter view, although they realized that if silica were leached from metamorphic iron formation, soft ore would be formed, and they must have observed such soft ore bodies, for they state: "The leaching of silica from this rock (amphibole-magnetite rock) would produce an ore as rich as that derived from the alteration of cherts, but as a matter of fact the silica is usually not leached from these rocks and ore deposits derived from them are small and rare" (1911, p. 553). They were impressed with the fact that the ores were associated with ferruginous cherts and slates, and, since they believed that these rocks were derived almost entirely directly from the original iron formation, it was perhaps natural to conclude that the metamorphic facies was inhospitable to ore formation. Gruner (1932, p. 191–192) in contrast states that, "The Vermilion ores were formed directly from a highly metamorphosed quartz - magnetite - hematite-amphibole rock, as proved by the highly folded and banded portions of jaspylite included in the porous ore." However Gruner (1946, p. 35, 98, 101, 112) seems to agree with Van Hise and Leith that on the Mesabi range the solution of chert was necessary for ore formation.

Goethite and hematite occur in the iron formations as residual products of the oxidation and leaching of a variety of primary and secondary iron silicates. Whether the alteration has taken place on a scale extensive enough to form the large ore bodies of the Lake Superior region may be questioned. Likewise the theory that the ores developed directly from original iron formation through the oxidation of siderite and leaching of the chert is based upon certain observed facts: Chert, siderite, and greenalite are present in the original iron formations, but are rare or absent in the hematite or goethite ores. The conclusion reached is that the ferrous minerals were oxidized and the chert re-

moved through solution. The removal of silica is necessary to form ore from iron formation, but that the silica was in the form of chert at the time of oxidation and leaching is not axiomatic. The silica may have been largely combined with iron in the form of iron silicates.

It is probably not generally appreciated that metamorphic iron formation is more abundant in some of the Lake Superior iron districts than either the oxidized or siderite-chert facies. For instance, in the Marquette district the aggregate surface area of the metamorphic facies is roughly 67 per cent of the total area of the iron formation, the siderite-chert facies about 3 per cent, and the oxidized facies about 30 per cent. The metamorphic facies accounts for about 60 per cent of the surface area of the iron formation on the Gogebic range—the remainder is oxidized. The siderite-chert facies does not occur in any large area in the Gogebic district, but is present locally in both the metamorphic and oxidized areas. Published information is inadequate to make similar estimates for the Mesabi range.

This data, although only approximate, indicates that the iron formation in the Marquette and Gogebic districts is largely metamorphic with only local areas of oxidation. Looked at in a broad way, there is nothing to indicate that the oxidized facies and associated ores were derived directly from the siderite-chert facies of the iron formation. In fact, when scale is considered as well as detail, the oxidized iron formation in general and the ores especially are best explained as derived from metamorphic iron formation.

The term "soft ore" is used in this paper as employed by the mining companies in the Lake Superior region. The term includes ores of a diversified mineral composition which are highly porous and very friable. Soft red hematite and martite are usually the chief iron minerals with specularite and goethite of secondary importance. Any one of the iron minerals may predominate in a given ore body or at a specific stratigraphic horizon with the other minerals occurring in lesser amounts or being absent. Pyrite, gypsum, apatite, kaolinite, and other minerals occur sporadically in relatively minor amounts. Chert or quartz is present in some ores and absent in others. The silica content of an ore as determined by chemical analysis does not serve to indicate the presence or absence of quartz.

TABLE 1.—*Total iron, silica, and alumina of ores which contain no quartz*

	Fe	SiO_2	Al_2O_3		Fe	SiO_2	Al_2O_3
Marquette 1	60.41	5.58	2.91	Gogebic 1	67.69	1.56	.69
Marquette 2	62.01	5.50	.64	Gogebic 2	60.56	6.32	3.67
Marquette 3	59.71	6.06	3.65	Gogebic 3	63.63	5.81	2.56
Marquette 4	57.41	7.32	4.77	Gogebic 4	67.91	1.40	.78
Marquette 5	59.44	6.36	4.04	Mesabi 1	58.64	5.99	2.87
Marquette 6	57.99	7.34	4.85	Mesabi 2	62.25	1.50	.65
				Mesabi 3	56.64	10.86	1.89

Table 1 gives the total iron, silica, and alumina of ores which contain no quartz or chert. A portion of each sample was dissolved in hydrochloric acid and the residue examined for quartz with the microscope. That the silica and alumina are present in the clay minerals and iron silicates is substantiated by the fact that if a piece of ore is allowed to remain in hydrochloric acid until the iron is entirely dissolved, the original form, shape, and size of the piece of ore is preserved in astonishing detail by gelatinous silica and alumina. The material may then be sectioned with a razor blade and examined under the microscope. Detailed structures of the original fragment may then be observed.

The following discussion is confined to the Marquette, Gogebic, and Mesabi ranges. The author believes that the conclusions reached may be applied, perhaps with certain changes in emphasis, to the other ranges of the Lake Superior region.

Occurrence and Origin of the Ore Bodies on the Marquette Range

General statement.—The ore bodies of the Marquett district classified on the basis of occurrence include, ore bodies in the Negaunee formation which occur, 1) on or near the footwall, 2) on top of sill, 3) as chimneys, 4) at or near the Goodrich conglomerate; and, 5) ore bodies in the Bijiki iron formation. Mr. R.M. Crump will present a paper on the orgin of the hard ores, which fall largely in group 4, in the near future.

Footwall ore bodies.—Many of the larger soft ore bodies of the district occur in the thinly bedded lower 300-400 feet of the iron formation, which is known locally as the "poker chip horizon." Although individual ore bodies vary somewhat in detail, the general features of the group are remarkably similar. The ore is usually adjacent to steeply dipping diked-faults and often extends up on the dike for some distances making the thickest part of the ore body. The dike if highly chloritized may be altered to a very soft greasy ore. The ore bodies thin out rapidly, away from the dikes, and tend to follow stratigraphic horizons but extend for much greater distances parallel to the dikes. The ore may be directly underlain by either the Siamo slate or the ferruginous cherts and slates of the iron formation. Zones of iron formation, which occur interbedded in the upper part of the Siamo slate, are sometimes altered to ore (Fig. 1). The chloritized slate and the interbedded iron formation are often altered to ore close to the dike, but away from the dike the ore tends to follow specific horizons in the iron formation. Ther seems to be little tendency for the Siamo slate to alter to ore away from the dikes. Quartz veins, aome of which are tourmaline-bearing, are frequently encountered in the ore.

Many of the ore bodies exist in structural troughs produced by faulting, but some occur on the upthrown side of diked faults. There seems to be no direct relationship between the amount of movement on the diked faults and the occurrence of ore; faults with little if any vertical displacement carry ore, whereas others with up to 1000 feet of vertical displacement are known to be barren.

Certain associations of the ore are rather constant. The larger ore bodies are associated with the thicker and more persistent dikes or occur where thin dikes are numerous. A similar relationship has been noted on the Vermilion range; Gruner (1926a, p. 639) quotes Wolff to

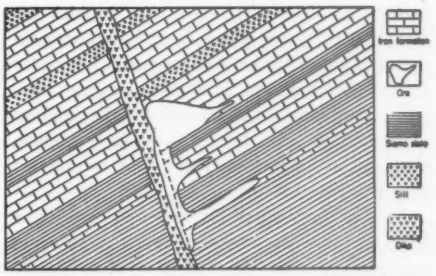

FIGURE 1.—CROSS SECTION OF AN ORE BODY IN THE TRACY TROUGH
The main portion of the ore body is developed in the basal beds of the Negaunee formation, but both dike and Siamo slate are locally altered to ore.

the effect that "That greater the intrusive the larger the ore bodies." Vertical dikes tend to have either ore bodies on both sides or no ore at all; inclined dikes (even with a steep dip) have the ore usually on the upper or hanging wall side.

The ores are characterized by high porosity and the ease with which they break down to a fine earthy mud. The dominant mineral constituents are soft red hematite with a varying, but usually subordinate proportion of martite. Goethite, magnetite, and specular hematite are present in lesser amounts, and gypsum is sporadic in its occurrence. The bedding is well preserved in the ore and usually is emphasized by concentrations of martite.

The footwall ore bodies, except for the Mary Charlotte, either lie within an envelope of oxidized iron formation, or have not been followed down the pitch far enough to determine the character of the limiting rocks. The derivation of the ores from metamorphosed iron formation is indicated by the widespread occurrence and abundance of martite and the location of ore bodies on the hanging wall side of dikes. The association of the larger ore bodies with the larger dikes, or their location in areas where dikes are more numerous, agrees well with the

concept. Furthermore, chloritized dikes and portions of chloritized Siamo slate have been converted locally to soft, porous ore.

The traditional view that oxidizing and leaching solutions, controlled throughout their course by impermeable horizons, have removed the chert from the original iron formation is untenable. The abundance of martite indicates that the ore was derived from the metamorphic facies of the iron formation; secondly the ores often do not lie upon impervious horizons; they may be enveloped by the "impervious horizon", i.e. lenses of ore in the Siamo slate, or the "impervious rock" (dikes and slate) may be altered to ore.

At the Mary Charlotte mine the gradation from ore to a metamorphic facies of the iron formation may be observed (Fig. 2). The ore body, which is of the stratigraphic type, passes down the pitch into a rock composed of partially oxidized stilpnomelane and siderite with subordinate chert, and then into an unoxidized stilpnomelane-siderite-chert rock. The Mary Charlotte ore body certainly appears to have been derived through the oxidation and leaching of iron formation composed dominantly of stilpnomelane.

Ore bodies on top of sills.—The ore bodies belonging to this class occur within the Negaunee formation, in some instances more than 1000 feet above its base. They assume the shape of great blanket deposits, lying on top of diabase sills. Large ore bodies such as the Lake Angeline, Salisbury, Lake Superior Hematite, and the Cleveland Lake, all south of Ishpeming, have been exhausted, but smaller ones still exist. The stratigraphy is not well enough known to determine whether these deposits lie upon one sill or several sills, but the latter seems more likely. The diabase sheets are offset by steeply-dipping diked faults with a general east-west trend. The vertical movement on the faults amounts in some cases to several hundreds of feet, with the south side moving upward in respect to the north side. This faulting has offset the southward-dipping sequence so that the same diabase sheet is brought to the surface repeatedly, thus forming the structural troughs to which Van Hise and Leith (1911, p. 271) refer.

From a mineral or textural viewpoint the ores are similar to those occurring at the foot-wall horizon. They are soft and porous, with the soft red variety of hematite usually the dominant mineral. Goethite is very abundant

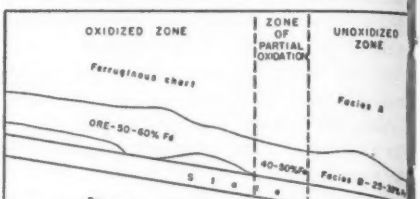

FIGURE 2.—CROSS SECTION SHOWING THE RELATIONSHIP OF THE FERRUGINOUS CHERTS AND ORE TO THE SILICATE FACIES AT THE MARY CHARLOTTE MINE

Facies A—Stilpnomelane with some siderite and magnetite and considerable chert.

Facies B—Stilpnomelane with subordinate chert, siderite, and magnetite.

at some localities. Martite is ubiquitous in its occurrence, and locally forms a blue, granular variety of ore.

The ore, according to Van Hise and Leith (1911, p. 462–475) was formed directly from the original iron formation by downward-circulating waters which oxidized the siderite and leached the chert. The structural troughs, bottomed by impervious igneous rock, concentrated the flow of the waters and thus localized the ore bodies. The field occurrence, however, indicates that the iron formation above the diabase sheets consisted largely of secondary iron silicates at the time it was oxidized and leached. Critical evidence bearing on this problem can be most readily observed at the Mitchell and Lothian properties south of Ishpeming. The principal ore body at the Mitchell Mine was located in the iron formation, at the top of a thick diabase sill, which dips steeply to the north. The ore body came in contact with the sill in places, but at many points oxidized iron formation intervened. Several ore bodies occurred as lenses within the upper part of the diabase. These lenses were of two types

bodies of ore lying entirely within the diabase sheet, and fingers of the main ore body which extended up the dip and were surrounded by the diabase on three sides. Relic metamorphic textures are common in the ore at this locality. Chert which could not combine with iron has recrystallized to coarse-grained quartz, and goethite and hematite often preserve the form of the iron silicates. If the iron formation is traced westward a few hundred feet to the Lothian property, the same sequence appears in the unoxidized state on an eastwark-facing cliff. The iron formation immediately above and included in the upper part of the sill is composed largely of grunerite and stilpnomelane. The partially oxidized and leached phase, transitional from the ore to the metamorphic iron formation is well represented upon the mine dumps of this area. At the Mitchell-Lothian mines, the conclusion seems inescapable that the soft ores were derived from metamorphic Negaunee iron formation.

In the Marquette district, ore bodies of this group usually occur on the top of thick diabase sheets. Thin sheets of basic igneous rock which may be tuff beds, lava flows, or sills occur at numerous horizons in the Negaunee iron formation, but they usually do not carry ore, even though perfect troughs were formed at dike intersections. The thin sheets certainly controlled the circulation of solutions to the same extent as the thicker ones, for oxidation extends into the bodies only a short distance from the upper or lower surfaces. The areas composed dominantly of iron silicates in the metamorphic iron formation bear the same relationship to the diabase sheets as do the ore bodies in the oxidized iron formation. Both are best developed on the top of the thicker diabase sheets, suggesting a direct genetic relationship.

Chimney ore bodies.—Only a few occurrences of this type are known in the Negaunee iron formation, but more will probably be discovered in the future. The Foley ore body at the Breitung Hematite mine is a typical example. It is a vertical chimney of ore, ranging from 200 to 300 feet in diameter. It has been mined downward from the surface for 1000 feet. The bottom of the ore body, as developed, is about 1000 feet above the Siamo slate; whether it extends to the slate footwall is not known. The Foley ore body is located at the intersection of east-west and north-south trending dikes, which should present a situation similar to the occurrence of closely spaced dikes. The combined metamorphic effect of the dikes might be expected to completely eliminate the chert through the formation of iron silicates and thus lead to an ore body after oxidization and leaching.

Ore bodies in the Bijiki iron formation.—The Bijiki iron formation, in the western portion of the Marquette district, has been severely metamorphosed by basic intrusives and the Republic granite. Several rather low-grade goethite ore bodies have been mined at various times, but the operations were marginal from an economic viewpoint. None of these mines is operating now. The ore bodies are located in a coarsely crystalline, metamorphic iron formation. The ores have been derived from a grunerite-magnetite rock which differed from the rest of the formation only in that sugary quartz was subordinate or entirely lacking. This relationship is readily observed, for all the structures and textures occurring in the grunerite-magnetite rock, such as radial and sheaf-like growths of grunerite, are preserved by goethite in the ore. Specimens that show the complete transition form the unoxidized grunerite-magnetite iron formation through the oxidized stage to ore may be obtained at the Marine, Northhampton, Phoenix, Pascoe, Hortense, and Bessie mines near Champion, or at the Titan, Beaufort, Ohio, Portland, Webster, or Imperial mines west of Lake Michigamme. The transition from the metamorphic iron formation to the ore at these mines clearly indicates that the ores were derived from the grunerite-magnetite rock.

Relatively small, low-grade ore bodies are typical of the oxidized and leached magnetite-grunerite phase of the iron formation. Large high-grade bodies of soft ore are not to be expected since the development of grunerite, as previously explained, does not bring complete elimination of the chert during metamorphism unless the original iron formation was excep-

tionally rich in iron. The remaining chert recrystallizes to coarse "sugary quartz", which is resistent to leaching.

Occurrence and Origin of Ore Bodies on the Gogebic Range

General statement.—The mode of occurrence of the ore bodies on the Gogebic range has been described by Van Hise and Leith (1911, p. 235–238) and Hotchkiss (1919, p. 25–27). Continuous mining and exploration has revealed additional details concerning the ore bodies but has added little to our understanding of the general occurrence. The ore bodies may be classified into: (1) trough-shaped ore bodies associated with dikes, (2) stratigraphic ore bodies, and (3) ore bodies associated with faults and fractures.

Trough-shaped ore bodies associated with dikes.—A prominent northeast-trending and a less well developed northwest-trending series of basic dikes dip southeast and southwest respectively. They intersect the northward-dipping Huronian sequence of Palms quartz-slate, Ironwood iron formation, and Tyler slate at about 90 degrees. The intersection of the northeast-trending dikes with the Palms quartz-slate footwall produces eastward-pitching troughs; the intersection of the northwest-trending dikes with the footwall produces westward-pitching troughs. Many of the larger ore bodies of the district have been found in these troughs. The ore bodies as seen in cross section usually have a southward-dipping limb that parallels the dike, as well as a northward-dipping limb that follows the footwall.

Van Hise and Leith (1911, p. 242–243) vigorously supported the theory that these troughs, composed of relatively impervious rock, were an important factor in the localization of the ores on the Gogebic range. The troughs were called upon to concentrate downward-moving surface waters that oxidized the siderite and leached the chert from the original iron formation. They believed that if a sufficient volume of water flowed through a given cross-sectional area of iron formation, the chert would be entirely removed and an ore body formed.

The derivation of ferruginous cherts and ore from the metamorphic facies of the iron formation does not require imprevious troughs. The occurrence of martite throughout the ore of the entire range and the gradation of ore into iron formation consisting dominantly of minnesotaite and stilpnomelane in sections 16, 17, and 18, 47-44 on the eastern Gogebic indicate that this ore was not derived directly from the original iron formation. How then is the occurrence of ore in the footwall troughs to be explained? It has already been pointed out that the elimination of the chert through the formation of secondary iron silicates is more complete above intrusives rather than below. Subsequent oxidation and leaching of the iron silicates form the ore bodies. The point should be emphasized that it is not a question of how much water flowed through a given cross-sectional area of the iron formation, but rather of the mineral composition of the area at the time of oxidation and leaching. Actually more solutions may have passed through brecciated cherty iron formation without producing any ore, than passed through an equal cross section of ore. It is interesting to note that both Hotchkiss (1919, p. 24) and Dickey (1929, p. 604) comment upon the very slight metamorphic effect of the dikes upon the oxidized iron formation and the ore. Oxidation and leaching tends to destroy the iron silicates, but in the unoxidized iron formation the abundance of iron silicates adjacent to dikes is very evident.

The shape as well as the location of the ore bodies is determined by factors which favored the development of secondary iron silicates from the chert and siderite. The more important factors are the hydrothermal metamorphic effect of the dike and the composition, texture, and permeability of the iron formation. Some dikes have undoubtedly exerted a greater effect upon the iron formation than others, either because they were intruded at higher temperatures or because hydrothermal solutions were present in greater quantities. One might expect that the ore bodies would tend to follow these dikes and in so doing cut across the bedding of the iron formation, but at the same time the ore would also tend to follow stratigraphic horizons in the iron formation composed dominantly of iron silicates. The result is a V-

shaped or trough-like type of ore body. Ore bodies of this shape are not dependent upon impervious troughs; they are also found at the intersection of dikes with ordinary iron formation. The development of the secondary iron silicates, oxidation, and leaching to ore in any favorable bed that comes in contact with the dike will produce exactly the same type of V-shaped ore body, provided the ore extends upward upon the dike. If the ore does not follow the dike, a lense-shaped or stratigraphic ore body results.

North-south trending dikes occasionally occupy fault zones in the iron formation. These dikes are nearly vertical or dip steeply eastward. Ore bodies seem to terminate against the dikes in some cases, and in others continue as if the dike were not present. The ore may extend up on the dike for a considerable distance producing a chimney-shaped body which cuts across the bedding of the iron formation. These ore bodies seem to be very similar to the chimney ore bodies on the Marquett range and probably are of the same origin.

Stratigraphic ore bodies.—It has already been pointed out that the texture and composition of a given stratigraphic horizon may favor elimination of the chert through the formation of secondary iron silicates. Oxidation and leaching of a silicate horizon of this type should lead to the formation of a stratigraphic ore body. This is probably the reason for the frequency of this type of ore body in the lower part of the iron formation at mines, such as the Castile, Sunday Lake, Brotherton, Mikado, and Eureka.

Ore bodies associated with faults and fractures.—It has long been known that some of the ore bodies on this range are associated with faults and fractures in the iron formation. The transverse faults, which strike nearly perpendicular to the formation and dip steeply, seem to have been particularly important in localizing chimney-shaped ore bodies. Van Hise and Leith (1911, p. 242–243) and Hotchkiss (1919, p. 27–28) suggest that circulating ground waters moved freely along the fault zones oxidizing the ferrous minerals and leaching the chert from the iron formation.

The field evidence seems to indicate, however, that the fault zones were channelways at a much earlier time for hydrothermal solutions which promoted the formation of secondary iron silicates at the expense of the chert. This relationship is particularly clear in section 8, 47-44 on the eastern Gogebic, where portions of the iron formation, which consist almost entirely of secondary iron silicates, are closely associated with quartz-filled fractures. The quartz probably was deposited as a late hydrothermal phase of igneous activity. Dickey (1929, p. 606) mentions the presence of vein quartz in the Bourne fault at the Montreal mine, and Allen and Barrett (1915, p. 47) refer to the abundant white quartz veins at, and in the vicinity of the Presque Isle mine, which "... have rendered worthless in some instances concentrations of hematite that would otherwise be valuable iron ore." The Sunday Lake fault zone, when traced into the footwall greenstones, is marked by quartz veins and lenses. One of these lenses observed in outcrop is more than 100 feet thick.

Occurrence and Origin of the Ore Bodies on the Mesabi Range

General statement.—Many of the ore bodies on the Mesabi range are larger than those of the Marquette and Gogebic ranges, and basic intrusives associated with the ore are small and rare. Wolff (1917, p. 236) classifies the Mesabi ore bodies into three groups: fissure, trough, and flat-lying ore bodies. The fissure ore body is an incompletely developed trough ore body, therefore these two types may be grouped together for the purpose of discussing their origin.

Fissure and trough ore bodies.—Fissure ore bodies are, as the name indicates, narrow concentrations of ore along well developed, steeply-dipping sets of fractures in the iron formation. The fracturing is related by Wolff (1917, p. 247–250) to the warping of the relatively brittle iron formation into broad anticlines and synclines. The fissure ore bodies, through an increase in size and frequency, merge into the trough ore bodies, which occur in basin-like structures, surrounded by gently dipping taconite. The basin structure apparently is secondary, and has developed entirely by the compac-

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tion and slumping of the highly porous ores. Trough ore bodies range in size up to those which are a quarter mile wide, almost a mile long and 400 feet deep. Several of these ore bodies with somewhat different trends may unite to form very large concentrations of ore. The trend of the ore bodies seems to be almost invariably parallel to prominent sets of joints.

Gruner (1930, p. 853, 1937, p. 123, 1946, p. 111) has repeatedly pointed out that the ore bodies do not usually continue up the dip to the point where the beds crop out. Unaltered taconite often forms the up-dip limits of the ore bodies. The margins of the ore bodies are usually abrupt, with the transition from ore to taconite taking place in some cases within a few inches.

Leith (1903, p. 265–272) and Van Hise and Leith (1911, p. 186) concluded that the ore bodies were formed by downward-circulating ground waters, which oxidized the ferrous minerals and leached the chert from the original iron formation. The Pokegema quartzite and impervious layers in the iron formation, as well as various structures such as bedding, fractures, and broad shallow synclines, are supposed to have controlled and concentrated the flow of the downward-moving waters, thus localizing the position of the ore bodies.

Gruner (1937, p. 122–124) has presented evidence supporting the theory that meteoric waters heated by magmatic emanations were responsible for the removal of the chert. Gruner (1946, p. 99–100) recognized that stilpnomelane and minnesotaite were altered to goethite and hematite but did not emphasize the importance of the alteration to ore formation. He also apparently did not realize that stilpnomelane and minnesotaite were of metamorphic origin.

The ores of the Mesabi range appear to have formed in essentially the same manner as those of the Gogebic and Marquette ranges, i.e. by the oxidation and leaching of iron silicates from a phase of the metamorphic iron formation containing little or no chert. This conclusion is reached by the following evidence: the abundance of martite in the ores throughout the range, the occurrence of the metamorphic minerals, stilpnomelane and minnesotaite in the unoxidized taconite of the central and western Mesabi, and the gradation of ore into taconite composed of primary (greenalite) and secondary iron silicates. The evidence from the Marquette and Gogebic ranges indicates that the association of magnetite with minnesotaite and stilpnomelane is indicative of the metamorphic environment. The transition from silicate iron formation which contains little chert to ore can be seen at many localities, but is especially well exposed at the Graham mine on the eastern Mesabi. The area of oxidation and complete leaching has been less extensive here than the area composed dominantly of iron silicates. If the area of oxidation is more extensive, oxidized siliceous taconite usually forms sharp contacts with the ore body. These abrupt contacts reflect the sharp limits of the essentially chert-free iron silicate areas which are observed in the metamorphic facies of the iron formation. When an ore body passes into unoxidized silicate iron formation which contains little chert, a narrow gradational zone of partial oxidation and leaching is present. Chemical analyses show that the iron content of this zone usually ranges in the high forties and low fifties.

On the eastern Mesabi for a distance of about 14 miles, the iron formation has been extensively metamorphosed by the Duluth gabbro to a magnetite-grunerite rock. As the iron formation is traced westward away from the gabbro, grunerite disappears, minnesotaite and stilpnomelane appear with magnetite and are present hroughout the central and western portions of the range. The magnetite-grunerite rocks of the eastern Mesabi are unoxidized and unleached, which accounts for the absence of soft ore bodies in this area. The lack of extensive oxidation and leaching of the magnetite-grunerite phase cannot be correlated with the mineral composition or the grade of metamorphism attained, but appears to be dependent upon other factors, such as the source of the oxidizing solutions or the lack of adequate channelways.

The large rich ore bodies of the central and western Mesabi appear to have been formed from the minnesotaite-stilpnomelane phase of the iron formation. The formation of min-

nesotaite and stilpnomelane requires a greater proportion of silica per unit of iron than does grunerite. Therefore, in a rock of a given iron content, the development of minnesotaite and stilpnomelane eliminates a larger quantity of chert, than would be required by grunerite, which leaves much of the chert to recrystallize in the form of sugary quartz. Thus large high-grade ore bodies would not be likely to form from the magnetite-grunerite phase of the iron formation, even though it is extensively oxidized and leached. The ore bodies developed in this facies of the iron formation on the Marquette range are numerous but neither large nor high grade.

Since intrusives in the iron formation other than the Duluth gabbro are rare on the Mesabi range, contact metamorphism cannot be of more than local importance, although Gruner (1937, p. 123–124) states that the evidence indicates the "Existence of large Keweenawan intrusives under the iron formation as proved by diabase intrusives in the iron formation at least as far west as Keewatin." However, quartz veins which mark channelways followed by hydrothermal solutions are numerous in the iron formation. Leith (1903, p. 212) points out the abundance of quartz veins in the ore, and Gruner (1946, p. 70, 98, 108, 109) mentions many quartz veins, some of which are 15 feet wide and can be traced for a distance of 2000 feet. The quartz veins were formed prior to the ore, for they are shattered and broken by the subsidence related to the slumping of the ore. Gruner (1930, p. 856) says: "These veins are certainly of hydrothermal origin. Their relation to the ore bodies is obscure, though some of them form the walls along which slumping has taken place in the ore bodies." The clear-cut relationship of intense metamorphism to quartz veins on the Marquette and Gogebic ranges indicates that the formation of secondary iron silicates may be the factor which relates the quartz veins to the ore bodies.

Flat-lying ore bodies.—Flat-lying ore bodies are erosion remnants of trough ore bodies, or ore layers extending away from a trough ore body. These ore layers follow stratigraphic horizons. Their mode of formation was probably the same as the trough ore bodies, the only variable being that elimination of chert through the development of secondary iron silicates was confined to specific stratigraphic horizons. Gruner (1946, p. 106) discussing a flat-laying ore body, states:

> "It occurs directly under the Intermediate Slate or Paint Rock and follows the silicate taconite horizon for distances in excess of one mile in extreme cases like the Godfrey mine. The thickness does not exceed 20 to 30 feet. In plan such ore bodies are very irregular and one wonders how such complete leaching of just one bed was possible."

The significant fact appears to be that the ore body follows the silicate taconite horizon. The ore body probably represents a portion of that horizon, which was composed largely of minnesotaite and stilpnomelane, and was later subjected to oxidation and leaching solutions.

Many of the wash ore bodies of the western Mesabi are of the flat-lying type. They are composed of oxidized taconite in which the quartz grains and iron minerals are in a loose, granular to friable condition. Gruner (1937, p. 129) and Leith (1903, p. 211, 296) conclude that the leaching of chert along the grain boundaries has removed the bond between the grains, allowing the rock to disintegrate into sand.

Thin sections of unoxidized taconite grading into wash ore, examined during the present study, show that in some cases the individual quartz or chert grains have minnesotaite or stilpnomelane concentrated along their boundaries. When these iron silicates are oxidized, the bound between the quartz grains is destroyed and the material becomes very friable. The same type of alteration has been observed in the Siamo graywacke of the Marquette district. The original rock consisted of quartz and feldspar grains with iron silicates (largely chlorite) forming the matrix. When the iron silicates are oxidized to hematite or limonite, the sand grains are freed and the material becomes loose sand. Gruner (1946, p. 98) describes an occurrence, which appears to be similar, at the Biwabik mine on the Mesabi range. He states: "In the Biwabik pit where the ore rested on the Pokegama quartzite, a foot or two of the quartzite apparently has been attacked by the leaching solutions leaving

a sand of loose quartz grains which was troublesome in the removal of the overlying ore." It seems probable that the matrix of the sand grains in the unaltered quartzite consisted of iron silicates, which upon oxidation and leaching freed the individual quartz grains.

Source of the oxidizing and leaching solutions

General statement.—The lack of unanimity among geologists regarding the source and direction of movement of the solutions that oxidized the ferrous minerals and removed silica from the iron formation has already been mentioned. The proponents of both the descending cold-water and the ascending hot-water theories assume that chert was removed in the large quantities necessary for ore formation. This line of reasoning forced them to conclude that the distribution and position of the ore bodies was determined by concentrated flow of the oxidizing and leaching solutions along certain channels. The intersection of dikes and impervious horizons has often been cited as the prevailing structures which controlled the flow of underground waters, and thus the oxidation and leaching on the Gogebic and Marquette ranges.

According to the concept advanced in this paper, ore bodies are located where the iron formation consisted largely of iron silicates formed during metamorphism, which were subsequently oxidized and leached. The oxidizing and leaching solutions, as far as we know at present, can not be genetically related to the formation of the iron silicates. The field evidence indicates that oxidation and leaching are definitely later than the metamorphism, extensive areas have been thoroughly metamorphosed but are without a trace of oxidation or leaching. Thus, position alone of an ore body can not be used as evidence for the direction of movement of the oxidizing and leaching solutions. The ore bodies are located above the dikes on the Gogebic range because of the factors controlling complete reaction of the chert with the iron during metamorphism, not because unusually large volumes of water was concentrated there. Larger volumes of the oxidizing and leaching solutions may have passed through brecciated portions of the cherty iron formation for long periods of time without forming any ore.

Factors which controlled the path of the oxidizing and leaching solutions.—The oxidized iron formation provides little direct evidence regarding the primary control on the movement of the oxidizing and leaching solutions. Of course, as areas of iron formation consisting dominantly of iron silicates were oxidized and leached of silica, they became more permeable. This increased permeability gradually tended to concentrate the flow of the solutions into the areas altering to ore.

The factors which exerted the initial control over the movement of the oxidizing and leaching solutions can be studied best where the oxidized iron formation is transitional either to the original iron formation or to the metamorphic facies. Secondary structures such as faults and joints seem to have exercised initial control over the movement of the solutions with stratigraphic horizons assuming a secondary role. On the eastern part of the Gogebic range, oxidation is confined to fault and fracture zones and certain members of the iron formation intersected by these structural channels. The oxidation often diminishes as the stratigraphic horizons are followed away from the fault zone, indicating that bedding control is secondary. Fault and breccia zones located in areas of unoxidized iron formation often show no evidence of oxidation. Apparently oxidizing solutions were not always available to take advantage of these natural channelways.

Van Hise and Leith (1911, p. 474–475) have emphasized the effect of impervious horizons, such as dikes and slates, upon the movement of the altering solutions, but in regions of partial oxidation it is the most permeable zones that controlled their flow. There are many occurrences in the Marquette district (Tracy, Milwaukee Davis, and Mary Charlotte mines) where unoxidized iron formation lies between the oxidized phase and the underlying slate, sill, or dike.

Evidence favoring downward-circulating waters.—The case for downward circulating meteoric waters as the oxidation and leaching agents necessary for ore formation, is not greatly weakened by the concept that the ores were derived from iron formation composed dominantly of secondary iron silicates.

Van Hise and Leith (1911, p. 554) speak of the iron silicates as," . . . resistant amphiboles which do not yield readily to the surface waters." Yet Leith and Mead (1912, p. 90–102) have no difficulty in explaining the derivation of the Cuban iron ores from serpentine by surface weathering, or the alteration of greenalite to hematite on the Mesabi range. Even Gruner (1932b, p. 191) who has consistently opposed the straight weathering hypothesis states: "One of the greatest obstacles for the acceptance of the weathering hypothesis is the fact that most of the silica that was leached was in the form of quartz and not in the state of an indefinite, more or less hydrous form of silica or silicate."

The environment of surface weathering is characterized by several important chemical processes which include oxidation, hydration, and hydrolysis, carbonation and desilication. Leith and Mead (1915, p. 37–44, 264) have emphasized the unstable nature of the iron-bearing silicates in this environment. The iron becomes oxidized to form limonite or hematite, and the silica is lost concurrently through solution. These reactions take place on a vast scale in the zone of weathering and produce striking results when the original rock is composed dominantly of iron-bearing silicates, as in the case of the Cuban iron ores. The rather sharp transition from ore to the unoxidized metamorphic iron formation in the Lake Superior region suggests the similar abrupt transition described by Leith and Mead (1912, p. 97) between the serpentine rock and the ore in the Cuban deposits. Although this type of transition might not be anticipated, it seems to be common to surface weathering.

An optimum surface-weathering environment during Precambrian time may have led to widespread oxidation and desilication of the silicate facies of the metamorphic iron formation. Furthermore it does not seem inconsistent that some of the ore bodies originated thus, whereas others were oxidized and desilicated by hydrothermal solutions.

Evidence favoring hydrothermal waters.—Gruner (1930, p. 851–857, 1937, p. 122–124) has gathered evidence both favorable and opposed to the hydrothermal leaching theory as applied to the Mesabi range. Since Gruner did not consider that large ore bodies were derived from silicate taconite containing little chert, he includes some evidence which bears upon the development and distribution of the silicates and not upon oxidation and leaching—for instance, the shape of the ore bodies, the abrupt contact between ore and taconite, and the fact that the ore bodies do not extend up the dip to the surface. However, other evidence, such as the great depths to which ore is known to extend, complete oxidation of magnetite to martite on a vast scale, alteration and bleaching of the Virginia slate where it overlies an ore body, and the existence of basic intrusives under the iron formation, applies equally well to the oxidation and leaching of secondary iron silicates.

The main body of thoroughly oxidized iron formation on the Marquette and Gogebic ranges shows little evidence as to the source or the direction of movement of the oxidizing and leaching solutions. The great depths (over 3500 feet) to which oxidation is known to extend suggests but does not prove that hydrothermal solutions were the active agents.

The basic dikes and sills on the Marquette and Gogebic ranges were more resistant to the oxidizing and leaching solutions than the silicate or carbonate formation; thus they provide an alteration intensity record at the numerous points where they have been penetrated by diamond drilling. Sills which lie above ore bodies on the Marquette range usually show the most intensive oxidation on the lower side of the sill, directly above the ore body. There seems to be no tendency for the sills to be more extensively oxidized on the upper surface than the lower, even away from ore bodies. This suggests that the oxidizing and leaching solutions were rising, but does not prove that they were hydrothermal.

The chloritized phase of the basic dikes has been altered extensively to paint rock, and locally in the vicinity of ore bodies to a soft, porous, greasy ore. It is very apparent that the oxidizing and leaching solutions were able, at least in some instances, to penetrate, oxidize, and leach these impervious rocks. The character of the alteration is even more interesting.

Table 2 gives the iron, silica, and alumina content of four examples of fresh, chloritized dike from the Tracy ore body on the Marquette

range and of the same material altered partially or completely to ore. The samples were obtained from diamond drill core, and each pair of analyses represent material taken from the same piece of core. A comparison of the analyses of the fresh and altered rock shows that alumina and silica have been lost relative to iron. Furthermore the silica-alumina ratio of the unaltered rock is maintained in the partly altered phase and in the ore. Both the alumina and the silica apparently were removed simultaneously by the solutions which oxidized the chlorite. This type of alteration is certainly not characteristic of surface weathering, where silica is lost in relation to alumina.

TABLE 2.—*Analyses of fresh and altered chloritized dikes*

	Fe	$\frac{\text{Fresh}}{\text{Altered}} \times 100$*	Al_2O_3	$\frac{\text{Fresh}}{\text{Altered}} \times 100$*	SiO_2	$\frac{\text{Fresh}}{\text{Altered}} \times 100$*	$\frac{SiO_2}{Al_2O_3}$
1 Fresh	27.05		18.25		20.28		1.11
Altered	58.79	46.0	4.74	385.0	5.32	381.2	1.12
2 Fresh	28.95		18.94		23.20		1.18
Altered	56.99	50.8	5.76	328.8	7.48	310.2	1.17
3 Fresh	27.73		19.68		24.02		1.18
Altered	55.31	50.2	6.00	328.0	8.00	300.3	1.26
4 Fresh	23.10		18.98		24.39		1.28
Partly altered	42.96	53.8	12.87	147.5	15.76	154.8	1.23
Highly altered	62.66	68.5	2.85	451.5	3.50	450.3	1.22

* The number of grams of altered rock necessary to produce the amount of that constituent originally present in 100 grams of fresh rock.

Another indication that the oxidizing and leaching solutions were not entirely of surface origin is the pronounced bleaching and sericitization of the Siamo slate and graywacke beneath the footwall ore bodies of the Marquette range. Where the basal iron formation is unoxidized, the Siamo is either a chloritic slate or a graywacke composed of quartz and feldspar grains in a chloritic groundmass. As an area of oxidation is approached, the chlorite disappears and red hematite and abundant sericite take its place. The slate may be altered to a soft red ore beneath the thickest part of an ore body; a short distance away it may be bleached to a soft, white sericitic slate. It is evident that iron has migrated, but whether enough to enrich the overlying ore body materially is not known.

The dikes on the Gogebic range are extensively oxidized and often kaolinized adjacent to ore bodies, but highly altered dikes occur in oxidized iron formation (section 12, 47-45, and sections 16, 17, and 18, 47-44) on the eastern Gogebic without an accompanying ore body. Since kaolinization may result from the action of surface waters, as well as epithermal solutions, the alteration cannot be considered of diagnostic value.

Gruner (1946, p. 107) mentions that where ore has been found beneath the Virginia slate on the Mesabi range, "... the slate is decomposed to kaolinite and quartz, and is bleached to an almost white color." Solutions certainly passed through the slate, which is surprising since slate is usually considered very impervious to surface waters.

Some of the common mineral associations found in the iron formation suggest hydrothermal alteration rather than surface weathering. If the oxidizing solutions were entirely of surface origin, one would expect that siderite and pyrite would be unstable, iron silicates somewhat more stable, and magnetite still more stable. Yet disseminated pyrite occurs in the ferruginous cherts as a minor constituent and massive pyrite is intimately intergrown with martite at the Old Volunteer mine in the

Marquette district. Furthermore martite is often associated with fresh chlorite and with siderite. Wienert (1933, p. 70–74) ascribes the alteration of magnetite to martite in the presence of siderite to oxidizing surface waters, but this does not seem at all likely. Gruner (1930, p. 704–705) has pointed out that the abundance and wide distribution of martite in the oxidized iron formation and ore suggests hydrothermal alteration, since magnetite is one of the more stable minerals under oxidizing conditions. These mineral associations and occurrences are not local details, which may be brushed aside by sweeping generalizations. They indicate that hydrothermal solutions have permeated the rocks and are probably at least partially responsible for the oxidation and leaching of iron silicates to ore.

According to the modified theory of ore formation advanced by Gruner (1937, p. 121–130), one might expect to find evidence that hydrothermal solutions had altered the rocks in and adjacent to the main avenues of escape, whereas laterally the alteration would have all the characteristics of surface weathering. The concept advanced by the author, that the ores were derived from a metamorphic facies rather than directly from the original cherty iron formation, does not alter this picture materially, but introduces an additional step in ore formation.

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AREAS COVERED BY PUBLISHED GE

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GIC MAPS ON DIFFERENT SCALES

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1125-1132, 3 FIGS., 1 PL., JULY 1949

GEOLOGIC MAPPING IN THE UNITED STATES

By Leona Boardman

Abstract

An inventory of geologic mapping in the continental United States shows the areal extent of published geologic maps, differentiated by scales, and gives the results of calculations of the approximate number of square miles geologically mapped in each State on several scales, with separate figures for the maps published since 1919. Several graphs compare the progress of geologic mapping in the different States and in different countries.

Acknowledgment

The information contributed by Dr. H. M. A. Rice, of the Geological Survey of Canada, and Mr. M. W. Pangborn, Jr., and Miss Taisia Stadnichenko, of the U. S. Geological Survey, on geologic mapping in other countries is gratefully appreciated. I am indebted also to other members of the U. S. Geological Survey for their advice and criticism, particularly to Mr. William W. Rubey, who has furnished valued suggestions throughout the study.

A nearly completed inventory of published geologic maps of areas in the continental United States provides for the first time a graphic picture of the location and extent of geologically mapped areas. Some of the results of the study are shown on Plate 1. No attempt was made to evaluate the quality of the geologic mapping. Thus some of the maps included are of doubtful value; many of them show the bedrock but not the surface geology, which is all important to engineering problems; others show the surface geology but not the bedrock geology. Additional unpublished geologic maps are on file in the Federal and State geological surveys, universities, and oil and mining companies, but information about them has not yet been assembled, and many of the maps are not available to the public.

The detailed results of the study of published geologic mapping in the United States are being prepared and published by the U. S. Geological Survey as a series of State maps. Each published geologic map is outlined on these State base maps; an explanatory key gives the source and date of publication, author, and scale. Publication of these geologic map indexes began in July 1947, and ten are now available—for Colorado, Iowa, Kansas, Missouri, Montana, Nebraska, North Dakota, South Dakota, Utah, and Wyoming. The scale of the indexes is 1:750,000, or about 12 miles to the inch. Outline patterns in four colors indicate approximate scales of the geologic maps. Each color pattern represents all geologic maps published in a given scale range and provides a convenient grouping for the surprisingly large number of geologic maps published on off-standard scales. Thus, the scale range 1:38,400 to 1:76,800 includes maps published on the standard scale of approximately 1 mile to the inch plus others that are nearer that scale than any other standard scale.

The first geologic maps of areas in the United States were prepared and published at private expense, but geologic mapping was early recognized as an important public responsibility. The geologic map of Massachusetts (Hitchcock, 1832) was the first geologic map prepared at public expense. It was followed by others in increasing numbers. Four geological surveys (Schmeckebier, 1904; Smith, 1918; Merrill, 1924) authorized by Congress explored large areas in the western United States between 1867 and 1878. These official surveys accomplished much reconnaissance geologic mapping, but much of the time they were competing with one another with consequent wasted effort and funds. This led to the establishment of the U. S. Geological Survey in 1879, which was charged with responsibility to prepare a geologic map of the United States.

Systematic geologic mapping of United States began with the preparation of the geologic folios by the U. S. Geological Survey. The first

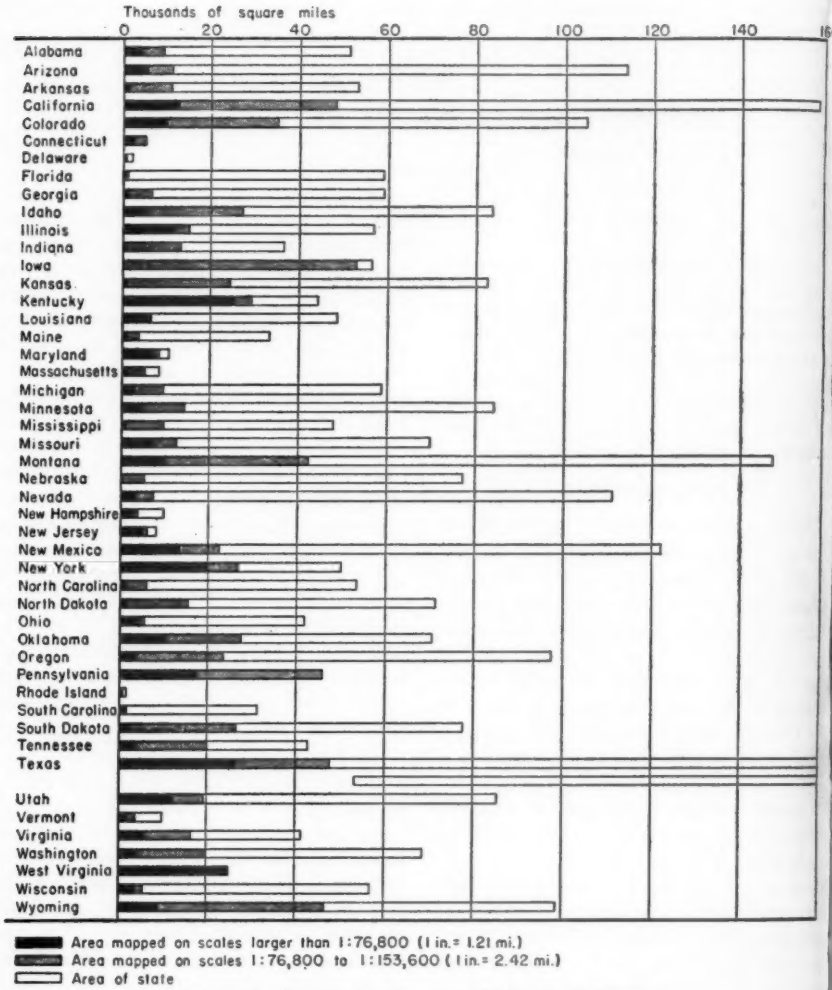

FIGURE 1.—PROGRESS OF GEOLOGIC MAPPING, BY STATES

folio was published in 1894, and the last of the series, No. 227, in 1945. Most of the geologic maps in the earliest folios and some of those in the later ones are 30-minute quadrangles on a scale of approximately 2 miles to the inch. The combined area of the folio maps on this scale is about 155,550 square miles. Most of the maps in the remaining folios are 15-minute quadrangles, scale approximately 1 mile to the inch; these cover roughly 26,900 square miles. Four of the folios contain geologic maps of a square degree on a scale of approximately 4 miles to the inch. The total area mapped in these 1-degree folios is about 14,700 square miles. Several other folios contain detailed geologic maps and descriptions of mining districts.

As the original geologic folio series was gradually discontinued, the Geological Survey published more geologic maps of irregular-shaped areas, as well as quadrangles, in its bulletins and professional papers. Because of the time required to prepare and publish these geologic

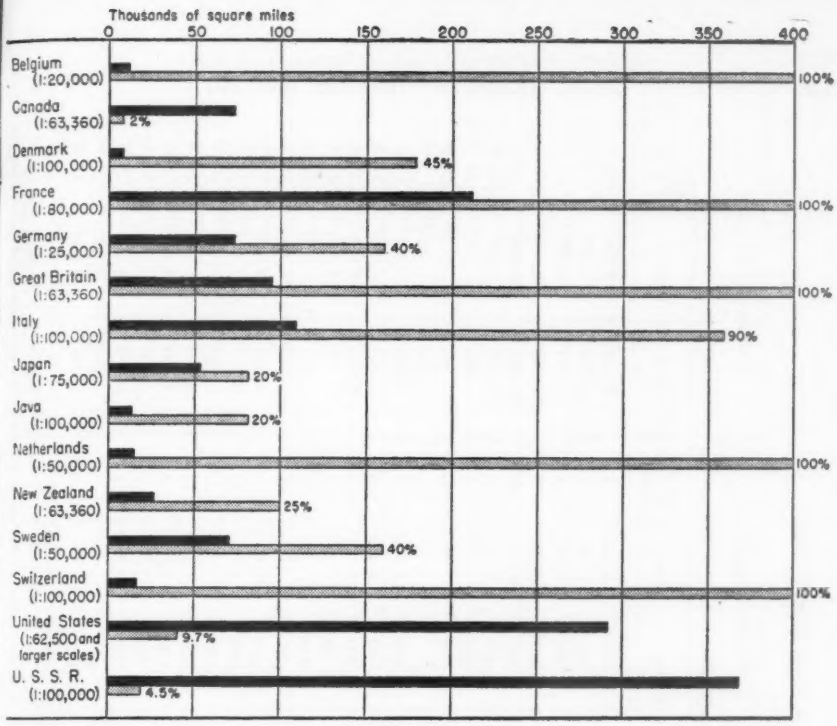

FIGURE 2.—COMPARATIVE PROGRESS OF GEOLOGIC MAPPING IN DIFFERENT COUNTRIES

Canada: Figures as of 1948 furnished by H. M. A. Rice, Office of Chief Geologist, Geological Survey of Canada.

United States: Approximate figures as of May 1948.

U. S. S. R.: Figures as of January 1, 1939, taken from *Mineralnaya syrievaya baza SSSR—Geologicheskaya izuchennost* (*Mineral raw resources base of the U.S.S.R.—The State of the geologic knowledge*), prepared for the 18th conference of the All-Union Communistic Party (bolshevik), M. L., 1939.

Other countries: Approximate percentages of countries mapped, as of 1940, furnished by Mr. M. W. Pangborn, Jr.

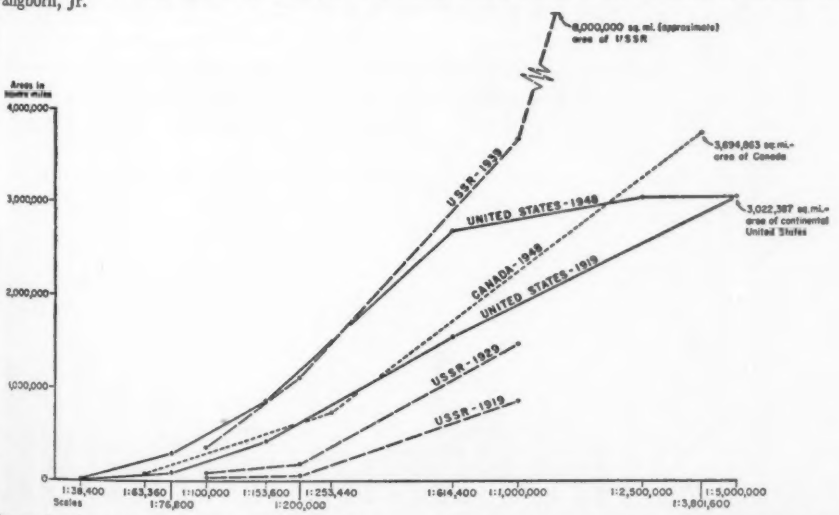

FIGURE 3.—AREAS OF PUBLISHED GEOLOGIC MAPS IN THE UNITED STATES, CANADA, AND THE U. S. S. R.

Points given represent number of square miles covered by geologic maps on the scales shown and all larger scales. *See* Figure 2 for sources of information on Canada and the U.S.S.R.

TABLE 1.—*Areas geologically mapped in the United States, by States**
(*Published maps only*)

State	Area of State† sq. mi.	1:38,400 and larger scales			1:76,800 and larger scales			1:153,600 and larger scales		
		1920–1948	Total		1920–1948	Total		1920–1948	Total	
		sq. mi.	sq. mi.	Per cent of State mapped	sq. mi.	sq. mi.	Per cent of State mapped	sq. mi.	sq. mi.	Per cent of State mapped
Alabama	51,609	250	320	0.6	4,400	4,500	8.7	4,500	9,000	17.4
Arizona	113,909	200	230	0.2	1,800	5,300	4.7	5,900	11,000	9.7
Arkansas	53,102	150	220	0.4	960	1,200	2.3	4,100	11,000	20.7
California	158,693	1,600	1,900	1.2	8,700	12,000	7.6	27,000	48,000	30.2
Colorado	104,247	840	1,400	1.3	6,300	10,000	9.6	18,000	35,000	33.6
Connecticut	5,009	18	25	0.5	1,600**	2,200	43.9	5,009	5,009	100.0††
Delaware	2,057	—	—	—	300	300	14.6	300	700	34.0
District of Columbia	69	50	50	72.5	50	69	100.0	50	69	100.0
Florida	58,560	—	—	—	—	—	—	1,200	1,200	2.0
Georgia	58,876	100	100	0.2	1,200	1,300	2.2	3,800	6,600	11.2
Idaho	83,557	180	190	0.2	4,700	5,200	6.2	23,000	27,000	32.3
Illinois	56,400	1,100	1,200	2.1	8,400	12,000	21.3	8,900	15,000	26.6
Indiana	36,291	28	57	0.2	28	450	1.2	28	13,000	35.8
Iowa	56,280	3	3	—	220	320	0.6	4,400	53,000	94.2
Kansas	82,276	190	190	0.2	580	750	0.9	18,000	24,000	29.2
Kentucky	40,395	1	3	—	24,000	25,000	61.9	24,000	29,000	71.8
Louisiana	48,523	4	6	—	5,900	6,400	13.2	5,900	6,400	13.2
Maine	33,215	12	13	—	150	760	2.3	1,900	3,400	10.2
Maryland	10,577	170	170	1.6	3,600	7,600	71.8	3,600	8,400	79.4
Massachusetts	8,257	150	180	2.2	1,600	1,800	21.8	3,300***	4,900	59.3
Michigan	58,216	500	990	1.7	910	3,100	5.3	2,200	9,400	16.1
Minnesota	84,068	410	640	0.8	1,300	4,100	4.9	5,000	14,000	16.6
Mississippi	47,716	22	22	—	800	800	1.7	9,000	9,500	19.9
Missouri	69,674	320	520	0.7	2,000	6,900	9.9	3,900	12,000	17.2
Montana	147,138	260	360	0.2	6,800	9,400	6.4	29,000	42,000	28.5
Nebraska	77,237	—	—	—	—	—	—	3,000	5,000	6.5

Missouri	69,674	320	320	0.7	2,000	6,900	9.9	3,900	12,000	17.2
Montana	147,138	260	360	0.2	6,800	9,400	6.4	29,000	42,000	28.5
Nebraska	77,237	—	—	—	—	—	—	3,000	5,000	6.5
Nevada	110,540	270	900	0.8	2,000	2,600	2.4	3,800	7,000	6.3
New Hampshire	9,304	11	11	0.1	2,500	2,500	26.9	3,200	3,400	36.5
New Jersey	7,836	15	130	1.7	1,800	4,500	57.4	1,800	5,700	72.7
New Mexico	121,666	470	490	0.4	12,000	13,000	10.7	18,000	22,000	18.1
New York	49,576	150	330	0.7	9,100	20,000	40.3	12,000	26,000	52.2
North Carolina	52,712	100	100	0.2	280	900	1.7	330	5,800	11.0
North Dakota	70,665	16	16	—	1,600	1,600	2.3	9,100	15,000	21.2
Ohio	41,222	—	—	—	3,000	3,400	8.2	3,600	5,200	12.6
Oklahoma	69,919	1,800	1,800	2.6	9,600	10,000	14.3	22,000	27,000	38.6
Oregon	96,981	240	250	0.3	3,100	3,300	3.4	20,000	23,000	23.7
Pennsylvania	45,333	50	1,300	2.9	8,500	17,000	37.5	8,800	45,333	100.0
Rhode Island	1,214	—	6	—	—	240	19.8	—	500	41.2
South Carolina	31,055	1	1	—	1,000	1,000	3.2	1,000	1,200	3.9
South Dakota	77,047	50	97	0.1	2,300	2,400	3.1	9,000	26,000	33.7
Tennessee	42,246	790	790	1.9	3,000	3,600	8.5	4,000	20,000	47.3
Texas	267,339	710	710	0.3	26,000	26,000	9.7	40,000	47,000	17.6
Utah	84,916	280	330	0.4	11,000	12,000	14.1	16,000	19,000	22.4
Vermont	9,609	120	220	2.4	1,200	1,500	15.6	2,100	3,400	35.4
Virginia	40,815	430	450	1.1	3,900	5,600	13.7	8,700	16,000	39.2
Washington	68,192	490	540	0.8	2,400	3,500	5.1	11,000	19,000	27.8
West Virginia	24,181	50	50	0.2	8,000	24,181	100.0	8,000	24,181	100.0
Wisconsin	56,154	110	370	0.7	110	3,400	6.1	180	4,900	8.7
Wyoming	97,914	430	610	0.6	6,300	8,300	8.5	23,000	46,000	47.0
		(13,141)	(18,290)		(204,988)	(291,970)		(440,597)	(816,192)	
Total	3,022,387	13,100	18,300	0.6	205,000	292,000	9.7	441,000	816,000	27.0

* Areas approximate and as of May 1948.

† From Sixteenth Census of the United States, 1940: U. S. Bureau of the Census, 1942.

** Maps for over three-fourths of this area show surface deposits only.

†† Glacial geology but not the bedrock geology of the State is mapped on 1:125,000 scale.

*** Maps of about half this area show only the Quaternary geology.

maps and their detailed text descriptions, the Survey has begun publication of a new series of geologic folios (maps with little or no text, to be followed by text descriptions later) to hasten the completion of the geologic mapping of the United States.

Several maps that show the general geology of the entire country on a small scale have been published. The most recent of these are the geologic maps of the United State (Stose, 1933), scale about 40 miles to the inch, and the geologic map of North America (Stose, 1945), scale about 80 miles to the inch. Geologic maps of 35 of the States[1] have been issued since 1900 on scales of 8 miles or less to the inch. A few of these maps were issued by the U. S. Geological Survey, but most of them were published by State Geological Surveys or other State organizations. Some of these State geologic maps are being revised, and geologic maps of several additional States are in course of preparation. The areas covered by these published State geologic maps, combined with areas geologically mapped in the other States on scales of 10 miles to the inch and larger, total approximately 2,700,000 square miles, or about 90 per cent of the area of the United States. Little more than one-fourth of the continental United States, or about 816,000 square miles, has been mapped on scales as large as 2½ miles to the inch; and geologic maps on scales as large as 1 mile to the inch have been published for slightly less than 10 per cent of the country, or about 292,000 square miles.

Larger-scale geologic maps of half a mile or less to the inch have been published for less than 1 per cent of the United States. The largest single area geologically mapped in a published report on scales more detailed than 1:38,400 is in the Osage Reservation in Oklahoma; about 1700 square miles in this reservation are geologically mapped on a scale of 1:31,250 (White and others, 1922). Between 400 and 500 square miles in the copper region of Michigan are mapped on a scale of 1:12,00 (Butler and others, 1929). Geologic boun aries have been added to aerial photograp of an area of over 400 square miles in the San Maria district, Santa Barbara County, Ca fornia, and published on a scale of 1:24,0 (Woodring and others, 1945). The coal fiel of King County, Washington, with an area more than 300 square miles, have been geolog ically mapped on a scale of 1:31,680 (Warre and others, 1945). About 20 additional area each larger than 100 square miles, have be geologically mapped on half a mile to the inc and larger scales.

Table 1 is based on estimates completed May 1948. Inasmuch as some of the are included in the inventory were mapped ma years ago, separate calculations were made show approximately how much of the tot geologic mapping has been done since 191 Figure 1 shows a graphic comparison of th progress of geologic mapping in the differen States on scales of about 1 mile or less to th inch, and additional mapping on scales of abou 2 miles to the inch.

During the period 1935–1939 roughly half the total area of about 39,000 square mil geologically mapped on a mile-to-the-inch an larger scales, and 60 per cent of the area of abou 71,000 square miles mapped on scales of abou 2 miles to the inch, were published by the U. S Geological Survey. The Federal Survey' share of this geologic mapping is probably some what more than these figures show becaus much of its geologic mapping is done in co-op eration with the State Geological Surveys which have published some of the maps pre pared by geologists of the Federal Survey Other geologic maps prepared by these geolo gists have been published in geological journal

Figure 2 shows the comparative progress o geologic mapping in the United States and othe countries. Whereas only about one-tenth o the United States has been geologically mappe on a scale of 1:62,500 or larger, geologic ma ping on comparable scales has been complete in several European countries. The proble of mapping the geology of this country is vast different, however, from that in any of the European countries, which are much small than the United States. On the other han

[1] Alabama, Arizona, Arkansas, California, Colorado, Connecticut (glacial), Delaware, Georgia, Idaho, Illinois, Indiana, Iowa, Kansas, Kentucky, Louisiana, Maryland, Massachusetts, Michigan, Minnesota, Mississippi, Missouri, Montana, New Jersey, New Mexico, New York, Ohio, Oklahoma, Pennsylvania, Rhode Island, Tennessee, Texas, Virginia, Washington, West Viginia, and Wyoming.

he United States with its large national income hould be expected to support a program of eologic mapping commensurate with its size. vailable figures on geologic mapping in the U. S. S. R. and Canada (Fig. 2), two other large ountries, show progress in geologic mapping ore nearly analogous to that in this country Fig. 3). The average annual rate of mapping, cale 1:100,000, in the U. S. S. R. in the 10-year eriod 1929–1939 is estimated to be about 8,500 square miles, compared to an average of 22,000 square miles a year in this country at miles to the inch and larger scales in the 5-year eriod 1935–1939.

The U. S. Geological Survey now is mapping pproximately 120,000 square miles, mostly a scales of 1 inch to the mile or larger. Figures re not available for geologic mapping in prog- ess by other organizations in this country.

A study of the progress of geologic mapping n any country suggests that there is no one "correct" scale and no ultimately complete eologic map of an area, but merely maps ufficiently complete and accurate for a given cale and purpose. The small-scale reconnais- ance map, usually prepared in advance of detailed mapping, has an important use in howing the broader geologic relationships. ubsequent geologic mapping, no matter on what scale, is almost certain to be followed by till larger-scale mapping with ever-increasing detail as required for special uses, both economic and scientific.

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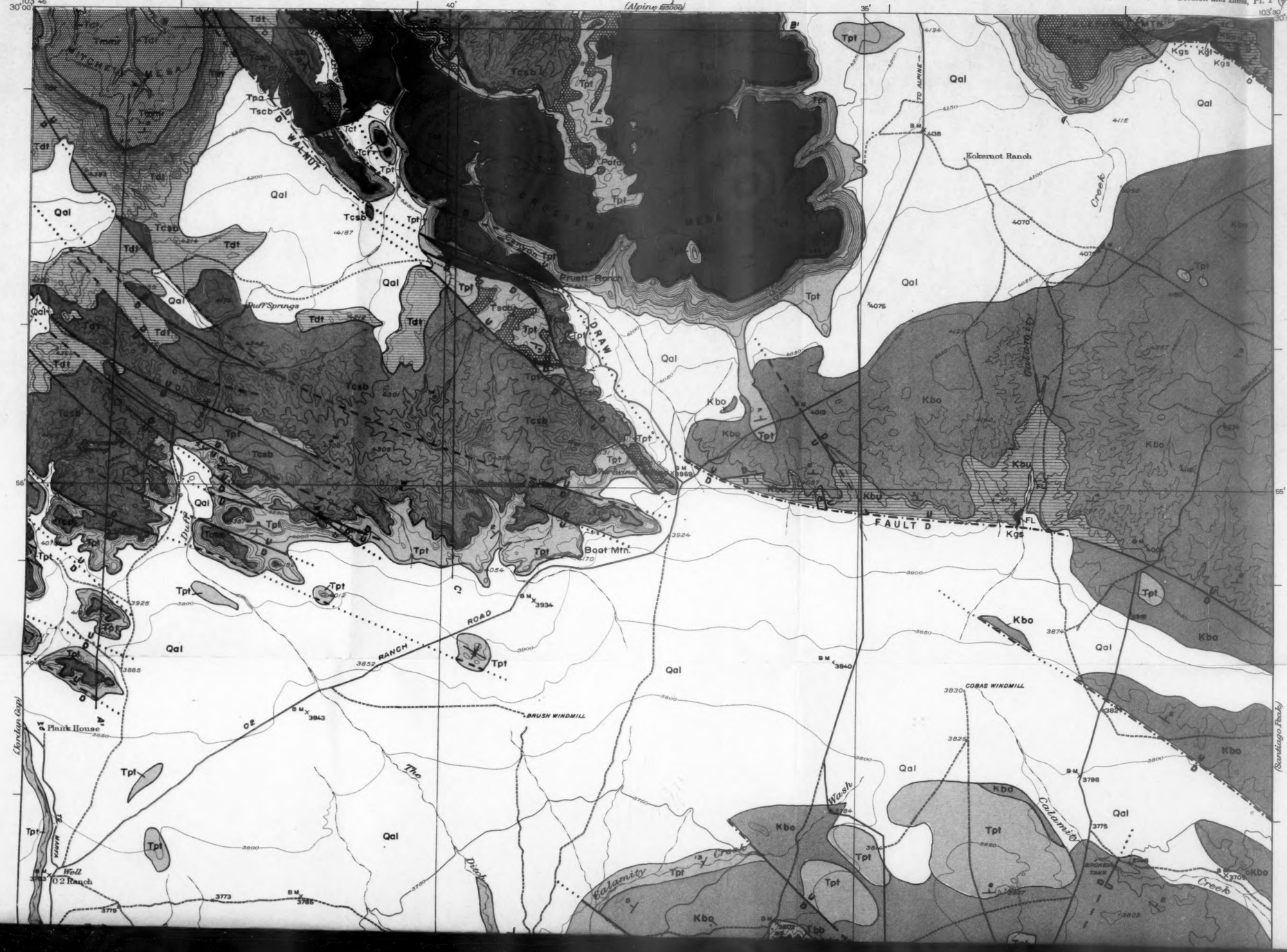


EXPLANATION

Qal — Alluvium

Intrusive igneous rocks

Syenite (Alkalic microsyenite)

Butcherknife basalt (Dense, black analcime trachybasalt)

Tmm — Mitchell Mesa rhyolite (Pink rhyolite porphyry flow)

Tdt — Duff Formation (Mainly light-colored rhyolitic tuff with minor breccia and dark-colored conglomerates)

Tcsb — Cottonwood Spring basalt (Dense to vesicular and amygdaloidal flows; trachyandesite to olivine basalt)

Tpt — Pruett formation (Tuff, breccia, conglomerate, sandstone, and fresh-water limestone with intercalated porphyritic flows: (Tpa) Potato Hill andesite; (Tscb) Sheep Canyon basalt; (Tct) Crossen trachyte)

UNCONFORMITY

Kbo — Boquillas formation (Alternating flaggy limestone and shale in lower part; thicker limestone and shale in upper part)

Kbu — Buda limestone (Massive limestone with middle marly unit)

Oligocene ? / Eocene ? — Buck Hill volcanic series

Upper Cretaceous

CRETACEOUS

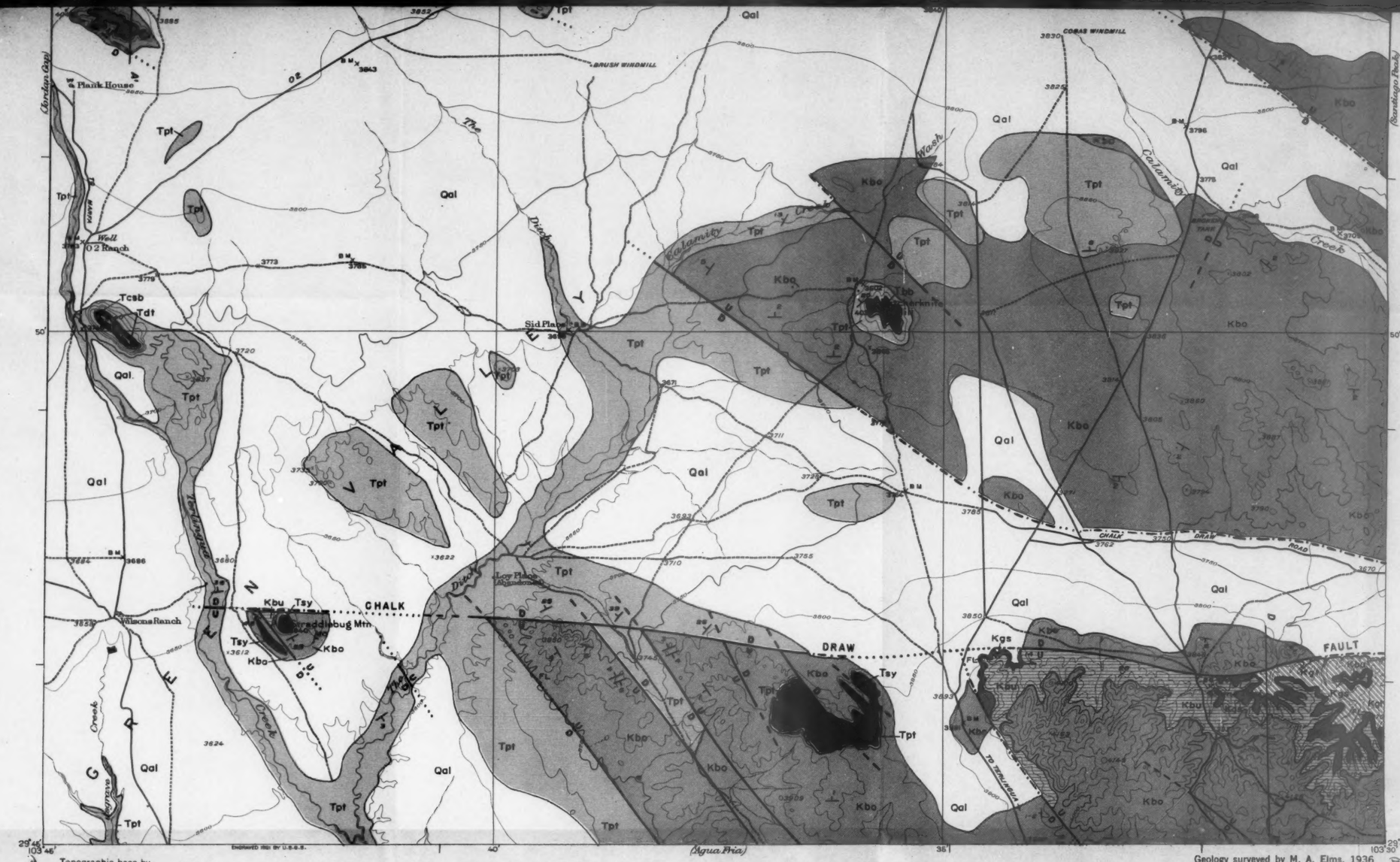
TERTIARY

QUATERNARY

CENOZOIC

GEOLOGIC MAP OF THE BUCK HILL QUADRANGLE, TEXAS

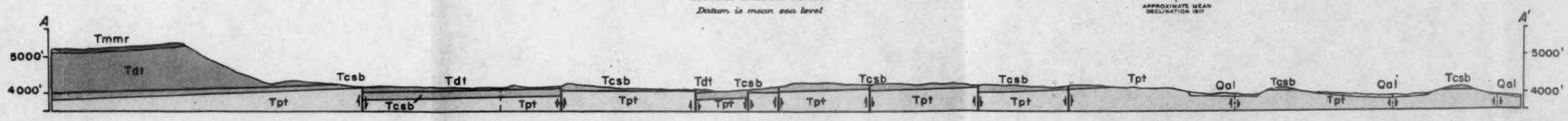
WITH STRUCTURE SECTIONS



Topographic base by
U. S. Geological Survey

Geology surveyed by M. A. Elms, 1936
and S. S. Goldich, 1940

SECTION A-A'



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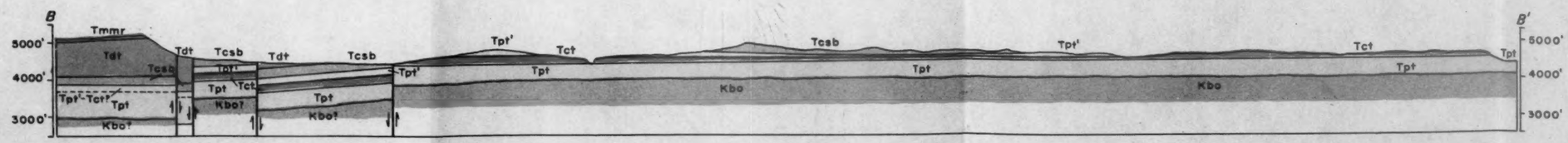
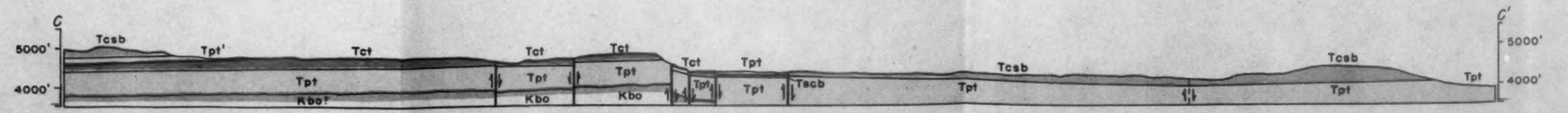


(Thin tuff beds and flows Tscb and Tpa included in Tpt')

SECTION C-C'



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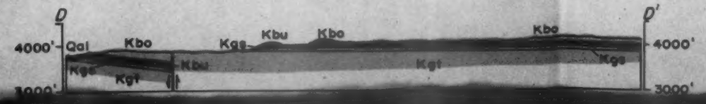

CRETACEOUS

Upper Cretaceous

Boquillas formation
(Alternating flaggy limestone and shale in lower part; thicker limestone and shale in upper part)

Buda limestone
(Massive limestone with middle marly unit)

Lower Cretaceous

Grayson clay
(Clay shale with thin calcareous sandstone layers)

Georgetown limestone
(Massive limestone; more thinly bedded in upper part)

and fresh-water limestone with intercalated porphyritic flows: (Tpa) Potato Hill andesite; (Tscb) Sheep Canyon basalt; (Tct) Crossen trachyte)

UNCONFORMITY

Kbo

Kbu

Kgt

U
D
Normal fault
U, *Upthrown side*
D, *Downthrown side*

Inferred normal fault

Normal fault concealed by alluvium; position marked by scarp

Normal fault concealed by alluvium

Strike and dip of beds

FL
Fossil locality

Water well

GEOLOGIC MAP OF THE BUCK HILL QUADRANGLE, TEXAS

WITH STRUCTURE SECTIONS

Topographic base by U. S. Geological Survey

Geology surveyed by M. A. Elms, 1936 and S. S. Goldich, 1940

ENGRAVED 1951 BY U.S.G.S.

Scale 1/62500

1 ½ 0 1 2 3 4 Miles

1 ½ 0 1 2 3 4 5 Kilometers

Contour interval 50 feet

Datum is mean sea level

APPROXIMATE MEAN DECLINATION 1917

Upper Cretaceous

Kbu — Buda limestone
(Massive limestone with middle marly unit)

Kdg — Grayson clay
(Clay shale with thin calcareous sandstone layers)

Kgt — Georgetown limestone
(Massive limestone; more thinly bedded in upper part)

Lower Cretaceous

CRETACEOUS

U / D — Normal fault
U, *Upthrown side*
D, *Downthrown side*

Inferred normal fault

Normal fault concealed by alluvium; position marked by scarp

Normal fault concealed by alluvium

Strike and dip of beds

FL — Fossil locality

Water well

SECTION A-A'

SECTION B-B'

(Thin tuff beds and flows Tscb and Tpa included in Tpt')

SECTION C-C'

SECTION D-D'

Scale of Sections

1000 0 5000 10000 15000 Feet

Abstr
Introd
Lo
Pr
Sc
Ac
Geog
Ph
Er
Cli
Creta
Ge
Ge
Gr
Bu
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Terti
Bu
Pr
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Figur
1.
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6.

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VOL. 60, PP. 1133-1182 6 FIGS., 6 PLS. JULY 1949

STRATIGRAPHY AND PETROLOGY OF THE BUCK HILL QUADRANGLE, TEXAS

By Samuel S. Goldich and Morris A. Elms

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Abstract

Tertiary tuff beds and related sediments with intercalated lava flows, aggregating over 2000 feet, are named the Buck Hill volcanic series for outcrops in the Buck Hill quadrangle, Brewster County, Texas. The volcanic series rests unconformably on the Upper Cretaceous Boquillas formation. The Lower Cretaceous is represented by the Georgetown limestone, the Grayson clay, and the Buda limestone.

The lower 900–1000 feet of the Buck Hill volcanic series is named the Pruett formation, and the upper 1000–1400 feet, the Duff tuff. Lava flows up to 325 feet thick, designated the Cottonwood Spring basalt, separate these formations. A limestone pebble conglomerate and arkosic sandstone at the base of the Pruett mark the Cretaceous-Tertiary unconformity. The Pruett tuff commonly is calcareous and grades to beds of fossiliferous freshwater limestone. A massive layer of breccia-conglomerate occurs near the top of the formation, approximately 60 feet below the base of the Cottonwood Spring basalt. Lava flows intercalated with the Pruett sediments, from oldest to youngest, are the Crossen trachyte, the Sheep Canyon basalt, and the Potato Hill andesite. The Duff formation is chiefly silicic tuff with minor breccia and a few thick beds of stream conglomerate. It is overlain by the Mitchell Mesa rhyolite, the youngest rock of the volcanic series in the quadrangle.

In the southern part of the quadrangle Straddlebug Mountain and Buck Hill are small intrusives of syenite; similar syenitic intrusives cut the Cretaceous, the Pruett tuff, and the Sheep Canyon basalt in the Elephant Mountain area. The igneous rocks are alkalic and chemically resemble the analyzed rocks from the Terlingua-Solitario region to the south.

The main structural features are west- and northwest-trending normal faults. Two major fault zones divide the quadrangle into three blocks with a middle down-dropped segment. Differential hardness of the rocks of the Buck Hill volcanic series controls the topography.

Introduction

Location

The Buck Hill quadrangle, west-central Brewster County, Trans-Pecos Texas, is on the southern front of the Davis Mountains (Figs. 1, 2). It is reached most easily from Alpine, 30 miles north, by the highway to Terlingua which traverses the quadrangle from north to south. Alternate routes are from Marathon by way of Del Norte Gap in the Del Norte-Santiago Mountain range and from Marfa by the 02 ranch road in Paradise Valley.

Previous Work

The older literature, including the pioneer work of Von Streeruwitz (1890), Vaughan (1900), Udden (1904; 1907b), and others, is reviewed in *The geology of Texas* (Plummer, 1932; Baker, 1934). The first petrographic work on the volcanic rocks of the Davis Mountains was by Osann (1892) on specimens collected by Von Streeruwitz. Lord (1900) made petrographic examinations of specimens collected by Vaughan in the Tierra Vieja Mountains. Much of the geologic knowledge of Trans-Pecos Texas is based on the extensive field work of C. L. Baker (1927; 1928; 1934; 1941; Baker and Bowman, 1917). King (1937) mapped the Marathon and Monument Spring quadrangles (Fig. 1). Eifler (1943) mapped the geology of the Santiago Peak quadrangle which adjoins the Buck Hill quadrangle on the east. Albritton and Bryan (1939) studied the Quaternary deposits of the Davis Mountains, including the

northern part of the Buck Hill quadrangle. Lonsdale (1940) studied the igneous rocks of the Terlingua-Solitario region in southern Brewster and Presidio counties.

Hardin's map of the Elephant Mountain area is reproduced in Plate 6. The base map is an enlargement of the U. S. Geological Survey topographic map of the Alpine sheet, a 30-min-

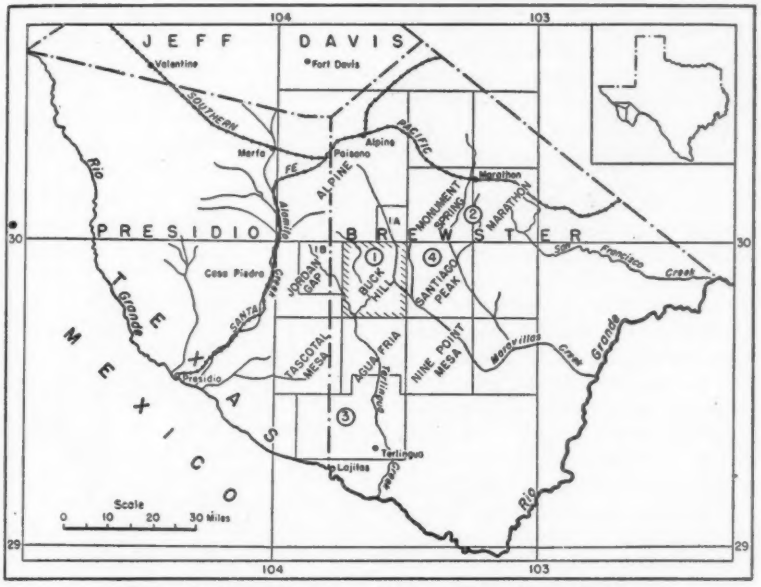

FIGURE 1.—INDEX MAP OF A PART OF TRANS-PECOS TEXAS

Showing Buck Hill quadrangle (1), Elephant Mountain area (1A), Paradise Valley area (1B), and related areas of published maps: (2) Monument Spring and Marathon quadrangles, King (1937); (3) Terlingua-Solitario region Lonsdale (1940); (4) Santiago Peak quadrangle, Eifler (1943).

Scope of Field Work

Good exposures of the Tertiary volcanic rocks and the underlying Cretaceous strata in the Buck Hill quadrangle make this area favorable for detailed studies of the Tertiary stratigraphy and petrology of the southern Davis Mountains. This investigation was started by Elms (1937) in the summer of 1936 and was continued and expanded by the writers in the summer of 1940. It became apparent that additional mapping and tracing of the rock units recognized in the Buck Hill quadrangle in adjacent areas was desirable. Reconnaissance mapping of the Elephant Mountain area in the southeastern part of the Alpine quadrangle was undertaken by Hardin (1942), and of the Paradise Valley area in the northeastern part of the Jordan Gap quadrangle by Graham (1942) during June and July 1941 under the supervision of Goldich. ute quadrangle (1:125,000) surveyed in 1893. The detail of this old map did not permit a close tie with the Buck Hill quadrangle, especially on the rugged slopes of Elephant Mountain. Additional field work was done by Goldlich in 1946.

Acknowledgments

The writers are indebted to Charles Laurence Baker for his stimulating interest and helpful criticism. Mr. J. E. Wilson assisted in the field work in 1936; R. W. Graves, Jr., and G. C. Hardin, Jr., in 1940; Hardin and D. W. Graham, in 1941; and C. L. Seward, in 1946. The paleontologic determinations contributed by a number of geologists indicated in the text are gratefully acknowledged.

A grant from the Penrose Bequest of The Geological Society of America defrayed expense

of a part of the field work and of most of the chemical work which was done in the Rock Analysis Laboratory of the University of Minnesota by arrangement with Frank F. Grout. This generous co-operation is appreciated.

Field work in 1946 and the preparation of the manuscript were done under the auspices of the Bureau of Economic Geology, The University of Texas, and valuable suggestions and criticism were received from staff members. This paper is one of a series reporting results of a Trans-Pecos Texas project sponsored by the Bureau of Economic Geology.

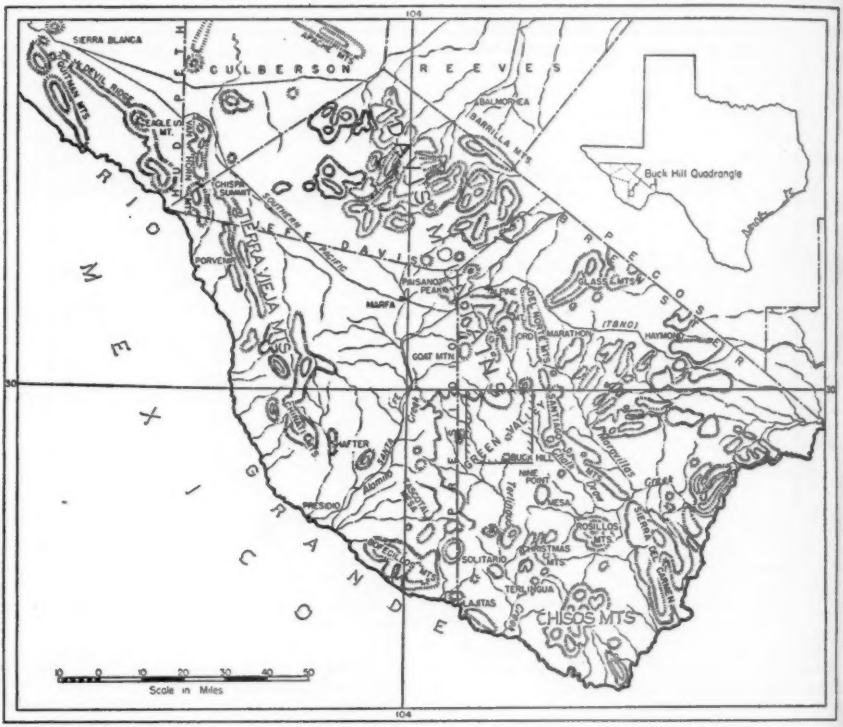

FIGURE 2.—SKETCH MAP OF A PART OF TRANS-PECOS TEXAS SHOWING PRINCIPAL PHYSIOGRAPHIC FEATURES AND LOCATION OF BUCK HILL QUADRANGLE

GEOGRAPHY

Physical Features

The Davis Mountains (Fig. 2) in northern Brewster and Presidio counties were carved out of a thick accumulation of volcanic tuff with numerous intercalated lava flows. The succession is pierced by many igneous masses. Late orogeny resulted in broad open folds in the Tertiary volcanics and in faults with a displacement of as much as 1000 feet, but the physiographic development of the region was largely controlled by differential erosion. Greater resistance of the flows and intrusives to erosion resulted in mesas and in prominent peaks.

Some of the most picturesque mesas of the southern front of the Davis Mountains extend into the northern part of the Buck Hill quadrangle. In the northeastern corner (Pl. 1) is the southern tip of Elephant Mountain which dominates the region with its great bulk and altitude of over 6000 feet above sea level. A massive caprock of microsyenite accounts for this large erosional remnant. West of Ele-

phant Mountain is Crossen Mesa capped by a thick flow of trachyte, and in the northwestern corner of the quadrangle is Mitchell Mesa with the southern tip of its rhyolite caprock rising 5351 feet above sea level. South of these mesas are lower ones capped with basalt. Their southern scarps face Green Valley which occupies the central and southwest part of the quadrangle. On the southeast, hills of Cretaceous limestone rise above the alluvium of Green Valley. The nearly horizontal Cretaceous strata continue eastward in the Santiago Peak quadrangle where they have been sharply upturned and thrust-faulted in the Santiago Mountain range.

Green Valley was formed during an earlier erosion cycle, and the alluvium of the valley now is being dissected and removed by the streams. The western part of the quadrangle is drained by Terlingua Creek, and the eastern part by the westernmost tributary of Maravillas Creek known here as Calamity Creek. The drainage divide in the quadrangle is poorly defined, and the surface water may pass from one drainage system into the other; yet the two streams join the Rio Grande 50 miles apart (Fig. 1). The streams have cut deep steep-sided channels in the alluvium of Green Valley and have renewed attack on the bedrock.

In the southern part of the quadrangle syenitic intrusives form Straddlebug Mountain and Buck Hill. In the southeastern corner prominent scarps are formed by the Cretaceous limestones.

Erosional Agencies

The chief erosional agency is running water. Although the annual rainfall is small, much of it comes as torrential downpours. Sheet wash is effective on the steep slopes. Undermining, sapping, and marginal fragmentation wear the more resistant rocks such as the igneous flows which cap the mesas. In this process the vertical joints, which permit downward seepage of water favoring weathering and spalling, are an important factor in the wasting and recession of the cliffs. Although the wind appears to blow constantly and occasionally reaches hurricane velocity, it is of secondary importance; the scanty vegetation of the region is much more effective against wind than against water erosion.

Climate and Vegetation

The Buck Hill quadrangle with an average annual rainfall of less than 15 inches is in the driest climatic province of Texas. Most of the rainfall comes during the hot summer months as thunder showers of short duration but often of great intensity and sometimes accompanied by violent winds. Temperatures during the day in the summer commonly are above 100°, but the nights are cool. The average mean annual temperature for Brewster County is approximately 63°.

The sparse vegetation includes a great variety of yuccas, cacti, and other hardy plants adapted to the semiarid climate. The relative abundance of certain plant types shows a response to physical factors. On valley flats sagebush and creosote are abundant. Dense growths of ocotillo characterize upland flats of volcanic tuff and breccia and are common on weathered slopes of andesite, basalt, and certain limestone formations. Rough hills and slopes have considerable sotol and lechuguilla. Ravines with better than average supply of water may be choked with catclaw, and occasionally cottonwood trees are found, as at Cottonwood Spring. Bunch grass is plentiful on the mesas but relatively scarce on many of the valley flats.

Cretaceous System

General Features

The Cretaceous formations (Table 1) represented in the Buck Hill quadrangle are the Georgetown limestone, of which only the upper 127 feet is exposed, the Grayson clay (Del Rio), the Buda limestone, and the Boquillas formation, which was eroded in part prior to the deposition of the Tertiary volcanics. The Cretaceous formations in this part of Trans-Pecos Texas are readily differentiated lithologically, but in comparison with north and south Texas the section is not well known, and there are many stratigraphic and paleontologic problems.

The Georgetown formation is a thick-bedded, commonly massive limestone. The Grayson is mainly clay and shale with a few thin calcareous sandstone beds near the top of the formation which contain abundant *Haplostiche texana* (Conrad), a large foraminifer. The Buda is chiefly thick-bedded, fine-grained, gray to white limestone with a nodular marly middle unit

TABLE 1.—*Geologic formations in the Buck Hill quadrangle*

Age	Group and formation		Thickness in feet	Correlation	Lithology and remarks
Quaternary		Alluvium	30 ±		Recent and older (Pleistocene?) unconsolidated silt and gravel.
		Angular unconformity			
(Oligocene?)	Buck Hill volcanic series	Mitchell Mesa rhyolite	38–150	Probably equivalent in part to Chisos beds, Vieja series, Square Peak volcanic series, etc.	Gray to pink rhyolite porphyry.
Tertiary		Duff tuff	1000–1400		Tuff, breccia, and conglomerate.
		Cottonwood Spring basalt	50–325+		Fine- to medium-grained basalt flows; dense to amygdaloidal; nonporphyritic.
(Eocene?)		Pruett tuff and intercalated flows: Crossen trachyte, Sheep Canyon basalt, and Potato Hill andesite	900–1000		Tuff, breccia, conglomerate, arkose, freshwater limestone; intercalated porphyritic flows of trachyte, basalt, and andesite.
		Angular unconformity			
Upper Cretaceous	Austin group	Undivided, included with Boquillas	210–300 (?)	Austin	Upper surface eroded. Limestone with calcareous shale near top.
	Eagle Ford group			Eagle Ford	Alternating shale and limestone flags in lower part.
		Disconformity			
Lower Cretaceous	Washita group	Buda	65–68	Buda	Thick-bedded limestone with middle nodular marly unit.
		Disconformity			
		Grayson	60–76	Grayson	Clay and shale with thin calcareous sandstone layers.
		Georgetown	127+	Georgetown	Massive, thick-bedded limestone; more thinly bedded in upper part.

which commonly is fossiliferous. The lower part of the Boquillas is typically thin argillaceous limestone flags separated by shale. The upper part is largely fine-grained, evenly bedded limestone with beds of marl or calcareous shale becoming thicker and more numerous near the top of the formation.

Georgetown Limestone

The larger of the two outcrops of the Georgetown formation in the Buck Hill quadrangle is in the southeastern corner (Pl. 1) just east of Stone Tank and continuing eastward in the Santiago Peak quadrangle. Here, along the Chalk Draw fault the Georgetown makes a prominent northward-facing cliff 100 feet high of massive limestone, except for the upper 38 feet which is distinctly bedded. Near Stone Tank a maximum exposed thickness of 127 feet of the Georgetown was measured. The outcrop is up to 1 mile wide, and the nonresistant Grayson formation has been stripped from most of its surface except for a small hill which is protected by a cap of the Buda limestone. The second exposure is on an east-west fault southeast of Elephant Mountain where a narrow strip of Georgetown limestone crops out.

The Georgetown limestone is gray to brownish gray and weathers to a dirty gray or white, rough, and etched surface. Potholes are common along creek beds, and the bare flat surfaces are covered with tinajitas or etched potholes (Udden, 1925). Beds in the lower part of the exposed formation are 10–15 feet thick, and these thick layers make the formation one of the most resistant in the area and form vertical cliffs and overhanging ledges. The upper 38 feet of the limestone is nodular, thinner-bedded, and fossiliferous. The texture ranges from fine- to coarse-grained. Fossil fragments which are numerous and scattered throughout the limestone are recognizable by their crystalline calcite. The lower 8 feet contains some chert in the vicinity of Stone Tank.

Only the upper contact of the Georgetown limestone is exposed, and near the base of Elephant Mountain the contact with the overlying Grayson is gradational. A 4-foot thickness of the Georgetown is overlain by nodular, marly, thin-bedded limestone, the "*Acanthoceras cunningloni* zone" (Adkins, 1932, p. 388), of the Grayson formation, which grades upward to a shell bed, 2–4 feet thick, composed almost entirely of the ramshorn oyster, *Exogyra arietina* Roemer, and of sandy marly clay in which this fossil is abundant. In the upper 50 feet of the Georgetown near Stone Tank, and in particular in the upper 38 feet of relatively thin-bedded limestone, fragmentary fossil remains are abundant, but few forms are identifiable. The brachiopod *Kingena wacoensis* (Roemer) is common in the upper 38 feet and extends about 15 feet lower. *Exogyra arietina* Roemer extends from the Grayson down about 8–10 feet in the Georgetown. One specimen of *Trigonia* sp. was found 8 feet below the top of the formation, and broken specimens of *Pecten* and *Gryphaea* were collected in the upper zone about 2 miles east of Stone Tank. The upper part of the Georgetown contains a Main Street fauna, but lack of identifiable fossils in the lower thick-bedded limestone precludes correlation with other faunal zones described by Adkins (1927, p. 50) in the Fort Stockton section where alternating limestone and marl beds of Georgetown equivalent have a total thickness of 245 feet. In southern Trans-Pecos Texas limestone replaces much of the marl of the Fort Stockton section, and the resultant massive reeflike limestone which is in part equivalent to the Georgetown formation (Washita group) and in part to older formations, Kiamichi and Edwards (Fredericksburg group), was named the Devils River limestone (Udden, 1907a). Recent work in areas east of the Buck Hill quadrangle shows that the fossiliferous gray to brown Kiamichi clay can be recognized, and the Devils River formation can be subdivided.

Grayson Clay

The largest outcrops of the Grayson (Del Rio) formation are in the southeastern part of the quadrangle. A good exposure is readily accessible just east of the Alpine-Terlingua highway south of the junction with the road to Marathon. The Grayson is 76 feet thick here and is capped with thick beds of the Buda limestone. East of Stone Tank the formation is about 70 feet thick, and its outcrop can be followed eastward into the Santiago Peak quadrangle. The beds dip less than 2° to the south. Southeast of Elephant Mountain the Grayson

is exposed in the scarp of a low southward-facing cuesta capped by Buda limestone. The outcrop is complicated by numerous faults and minor folds, but the average dip is about 3° to the northeast. In the channel of Calamity Creek 4½ miles south of the Kokernot ranch house there are small outcrops where much of the Grayson clay is eroded from beneath the Buda limestone leaving the latter as an overhanging ledge. Here the dip is approximately 4° N. The Grayson is about 60 feet thick in the northern part of the quadrangle.

The Grayson formation is dominantly clay-shale but also contains thin flags of limestone and sandstone. The clays usually are light buff, tan, or light greenish yellow, weathering to a characteristic tawny yellow with locally abundant secondary gypsum and limonite. The flags are most abundant near the top of the formation and are dark to light gray, usually very sandy, nodular and lenticular; a bed 6 inches thick may disappear laterally within a distance of 50 feet. The middle and lower parts are usually covered with slumped clay, so that fresh exposures are rare. The detailed section of the upper part measured in Calamity Creek (Appendix, Section No. 1) is typical of the formation in the central and southern parts of the quadrangle. At the foot of Elephant Mountain (Appendix, Section No. 2) and elsewhere in the northern part of the Buck Hill quadrangle, the upper part of the Grayson contains relatively more sand.

The contact of the Grayson with the underlying Georgetown rarely is exposed, but at Elephant Mountain the lower contact is conformable and gradational. The contact with the overlying Buda is an undulating surface with a sharp break from clay of the Grayson to nodular limestone of the Buda. In the Buck Hill quadrangle there are no marked indications of an unconformity, but elsewhere in Trans-Pecos Texas the Grayson formation is variable in thickness and locally absent. East in Val Verde County in road cuts on U. S. Highway 90, 1.3 miles east of Comstock, L. W. Stephenson found included in the lower beds of the Buda limestone cobbles and boulders derived from *Exogyra arietina* shell layers of the Grayson formation.

Fossils collected from the Grayson in the Buck Hill quadrangle are typical of the south Texas Grayson fauna. Three fossil localities are shown on the map: (1) near the base and southeast of Elephant Mountain, (2) in Calamity Creek, 4½ miles south of the Kokernot ranch house, (3) at the small hill (3970 feet) 1¾ miles east of Stone Tank. The ammonite *Turrilites brazoensis* Roemer occurs near the base of the formation. *Exogyra arietina* Roemer ranges 30 feet up from the base but was not found in the upper half of the formation. The arenaceous foraminifer *Haplostiche texana* (Conrad) is abundant throughout, especially in the sandy flagstones. About 2 feet below the top of the formation at the Elephant Mountain locality is a conspicuous shell bed, 4 inches thick, which contains abundant *Exogyra cartledgei* Böse, *Gryphaea graysonana* Stanton, *Exogyra whitneyi* Böse, *Ostrea* sp., and *Pecten* fragments. *Heteraster inflatus* (Cragin) is abundant in the upper part of the formation.

A large collection of dwarfed limonitic ammonites and echinoids was made at the Stone Tank locality, but this collection has not been studied. Most of the fossils listed were collected in either the lower 6 feet or upper 20 feet of the formation:

Haplostiche texana (Conrad)
Holectypus cf. *limitis* Böse
Heteraster inflatus (Cragin)
Protocardia texana (Conrad)
Inoceramus spp.
Pecten (*Neithea*) *budensis* Kniker
Exogyra arietina Roemer
cartledgei Böse
whitneyi Böse
Gryphaea graysonana Stanton
Ostrea sp.
"*Acanthoceras*" sp. aff. A. *cunningtoni* Böse
Turrilites brazoensis Roemer
Engonoceras sp.

Buda Limestone

The principal outcrops of the Buda limestone are along Calamity Creek north of the Walnut Draw fault and in the southeastern part of the quadrangle south of the Chalk Draw fault. The Buda also is exposed southeast of Elephant Mountain and on the northwest flank of Straddlebug Mountain in the southwestern part of the quadrangle where the Cretaceous rocks have been domed by a small syenite intrusive. Erosion has stripped the less resistant flags of the overlying Boquillas leaving wide flat surfaces of the white to gray fine-grained nodular Buda limestone. Commonly a heavier growth of oco-

tillo serves to differentiate the Buda outcrops from the Grayson and Boquillas. Although in places the Buda resembles the Georgetown limestone, its lighter color and more distinct bedding generally identify it.

The Buda of this area commonly has three lithologic facies: upper and lower units of thick-bedded, light-gray, fine-grained to porcelaneous limestone and a middle unit of light-gray to buff marl and thin-bedded nodular limestone. The upper and lower units with beds 4–6 feet thick form cliffs or steep slopes; the middle unit weathers to a gentle slope. At Elephant Mountain the Buda is 65 feet thick with a middle marly unit of 16 feet. In Calamity Creek the middle marly unit is 20 feet thick. Although the section (Appendix, Section No. 1) measured in Calamity Creek is incomplete, it is representative of the Buda lithology. Two miles east of Stone Tank the formation is 68 feet thick, and the marly unit measures 18 feet.

The Buda generally is fossiliferous with the middle marly unit affording the best material. Good fossil localities along Calamity Creek are indicated on the map (Pl. 1). The echinoid *Hemiaster calvini* Clark is abundant and marks a horizon about 20 feet above the base of the formation in the lower part of the middle marly unit. About 2 feet above the *Hemiaster* horizon is a 4-foot zone in which *Pecten* predominates. A third zone 2–4 feet higher is characterized by *Leiocidaris hemigranous* (Shumard). Ammonites were collected in all the zones, but these have not been studied in detail; *Budaiceras* spp. occur lowest in the section with *Hemiaster calvini* Clark.

Two fossil crab specimens from the middle marly unit were identified as *Graptocarcinus texanus* Roemer by Stenzel. The species was first described (Roemer, 1887) from the Buda limestone in Austin, Texas, and redescribed by Stenzel (1944). It had hitherto not been known outside of central Texas.

Fossils found in the Buda formation in the Buck Hill quadrangle are:

Hemiaster calvini Clark
Heteraster cf. *traski* (Whitney)
Leiocidaris hemigranosus (Shumard)
Protocardia texana (Conrad)
Salenia sp.
Pecten (*Neithea*) *texanus* Roemer
(*Neithea*) *whitneyi* Kniker
(*Neithea*) *subalpinus* Böse
(*Neithea*) *subalpinus* var. *linki* Kniker
Nautiloid sp.
Turrilites sp.
Budaiceras spp.
Graptocarcinus texanus Roemer

Boquillas Formation

All the strata above the Buda limestone and below the basal Tertiary conglomerate in the Buck Hill quadrangle are tentatively assigned to the Boquillas formation. Clayey and sandy limestone flags alternating with beds of shale are typical of the lower part of the formation. This succession is similar to the beds above the Buda limestone in southern Brewster County named by Udden (1907b, p. 29–32) the Boquillas flags. The upper beds are mainly limestone with some marl and shale which may be equivalent to part of Udden's (1907b, p. 33) Terlingua beds. How much of the Boquillas is Eagle Ford and how much Austin equivalent cannot be said; zoning or redefining of the Boquillas must await more detailed studies.

Large areas in the eastern part of the quadrangle are covered by the Boquillas formation (Pl. 1). South of the Chalk Draw fault the lower flaggy Boquillas caps the upthrown block east of the Terlingua highway. In the downthrown block the Boquillas beds dip south and rest against the Georgetown, the Grayson, and the Buda formations. West of the highway in the vicinity of Buck Hill the upper limestone beds cap the upland south of the Chalk Draw fault. North of Chalk Draw road an extensive outcrop of the Boquillas is cut off by a northwest-southeast fault west of Butcherknife Hill where the Pruett tuff is downfaulted against the limestone. This outcrop, 3–4 miles wide, extends eastward into the Santiago Peak quadrangle.

In the northeastern part of the quadrangle the Boquillas forms a southward-facing escarpment along the Walnut Draw fault extending from a point just east of Whirlwind Spring beyond the eastern boundary of the quadrangle. Most of the downthrown block south of the Boquillas escarpment is covered by alluvium, but east of Calamity Creek upper Boquillas limestone capped with a small remnant of the Preutt tuff is downthrown against the lower flags. The Boquillas also crops out southeast of Elephant Mountain, and the lower flags form a crescent-shaped outcrop around the syenite core of Straddlebug Mountain.

The gray to tan or brown limestone flags and interbedded shale in the lower part of the Boquillas formation are easily recognized. In thelower 50 feet, limestone flags, 2–4 inches thick, and shale beds, 3–4 feet thick, are evenly spaced, and because the shale is less resistant it erodes in a "stairstep" topography in which the flags form treads at 3- to 4-foot intervals. A sample of the flags dissolved in dilute hydrochloric acid gave an insoluble residue of about 14 per cent by weight. Most of the insoluble material is finely divided clay mineral of the montmorillonite group. The larger clastic grains are silicic plagioclase and microcline; quartz is rare. On many of the thinner flags there are markings which Udden (1907b, p. 32) described as fossil ice crystals.

The upper part of the Boquillas is mainly grayish-white to buff limestone weathering to light gray or white. The rock is fine-grained and uniform, locally with pyrite concretions which are commonly altered to limonite. The beds range from 1 inch to 4 feet thick, and the thicker beds of light-colored limestone differentiate the upper part of the Boquillas from the lower brown flaggy unit. There are some shale or marl breaks, however, and the distinction between lower and upper units on lithologic characters is not made easily everywhere. Calcareous shale in the upper part of the Boquillas beneath the basal Tertiary conglomerate is exposed just south of the Terlingua road ¼ mile southeast of the highway bridge on Calamity Creek Wash and at Hill 3927 approximately 2¼ miles southeast of the bridge and 2¼ miles northeast of Butcherknife Hill. The fresh shale is gray, weathering to a yellow brown or tawny yellow. At Hill 3927 the calcareous shale is 30 feet thick and resembles Terlingua beds along Terlingua Creek south of the Buck Hill region, but the outcrop is limited, and alluvium obscures the relations with the Boquillas limestone.

Because of pre-Pruett as well as recent erosion and because of faults and local thick cover of alluvium, a satisfactory measurement of the thickness of the Boquillas could not be made in the Buck Hill quadrangle. In the southern part of the quadrangle beds assigned to the Boquillas aggregate 210 feet. Near Stone Tank the lower flaggy beds are 130–140 feet thick. The measured section (Appendix, Section No. 3) southeast of Elephant Mountain includes 64 feet of typical lower flaggy beds and about 53 feet of upper limestone beds. The upper part of the section is covered, and probably 20 feet of additional strata beneath the Pruett tuff can be assigned to the Boquillas giving a total thickness of 127 feet. The formation probably is thickest in the hills east of Calamity Creek southeast of the Kokernot ranch house. Although no measurement was made, the maximum thickness is thought to be less than 300 feet.

The Boquillas and the underlying Buda in the Buck Hill quadrangle are concordant, but there is a marked lithologic change from the uppermost light-gray, thick limestone bed of the Buda to the lowermost dark-gray, sandy, banded, thin limestone flags of the Boquillas. Roberts (1918, p. 20) discusses the disconformable relations of the Boquillas to the Buda in Val Verde County. In road cuts along U. S. Highway 90 near Langtry the lowermost Eagle Ford bed is a sandy limestone. The contact with the underlying Buda, like that in the Buck Hill region, generally is undulating, but locally in these road cuts the basal Eagle Ford limestone is conglomeratic.

The upper contact of the Boquillas formation in the Buck Hill quadrangle is an erosion surface unconformably overlain by the basal conglomerate and conglomeratic sandstone of the Pruett formation. At some localities, as on either side of the O2 ranch road, 1½ miles northeast of Whirlwind Spring, the conglomeratic sandstone of the Pruett rests on limestone. At other places, such as Hill 3927, the basal conglomerate lies on stratigraphically higher beds of calcareous shale.

Identifiable fossils in the Boquillas are rare. Imprints of *Inoceramus* are abundant but cannot be identified specifically. The thin sandy flags, 12–20 feet above the base of the formation, contain a poorly preserved ammonite fauna which according to W. S. Adkins (personal communication, 1936) appears to be an unrecorded assemblage. Fragments of a fossil fish from the upper limestone were examined by the late Curtis J. Hesse (personal communication, 1937) who wrote:

"Your specimen is very incomplete, too much m for accurate determination. However, it appear to be similar to *Ferrifrons rugosus*, a fish describe

by D. S. Jordan (Univ. of Kansas Sci. Bull., vol. 15, No. 2, p. 228, pl. 18–19, 1924), from the Niobrara Chalk of Kansas. This, of course, is a somewhat meaningless name since Jordan could not tell what the fish was related to; but it is interesting that the two records should be of approximately the same stratigraphic position."

The calcareous shale below the basal Pruett conglomerate at Hill 3927, northeast of Butcherknife Hill, contains impressions of ammonites and pelecypods, but these are unidentifiable. Foraminifera were noted in a disintegrated and washed sample, and a study of the microfauna was generously undertaken by Dr. D. L. Frizzell and Miss Glenn G. Collins of the Department of Geology, The University of Texas. Frizzell (personal communication, 1947) wrote:

"Foraminifera are moderately common but in small variety, only seventeen species having been found in the sample. Although the following list of identifications is to a degree tentative, a correlation with the lower part of the Austin group seems definite.
Trochammina sp.
Lenticulina sp. aff. *L. münsteri* (Roemer)
Dentalina, 2 spp.
Gümbelina moremani Cushman
G. pseudotessera Cushman
G. sp. cf. *striata* (Ehrenberg)
Bulimina fabilis Cushman and Parker
B. sp. cf. *B. reussi* Morrow
Neobulimina irregularis Cushman and Parker
Valvulineria sp.
Globigerina sp.
Hastigerinella alexanderi Cushman
H. watersi Cushman
Globotruncana canaliculata (Reuss)
G. fornicata Plummer
Planulina austiniana Cushman"

Because of its stratigraphic position above the Buda limestone the lower part of the Boquillas formation in the Buck Hill quadrangle is correlated with the Eagle Ford; the uppermost part, from the fossil data now available, is lower Austin. Much remains to be done in the study of the fauna of the Boquillas in the Buck Hill quadrangle and in Trans-Pecos Texas. In this study the Forminifera will provide valuable information to supplement the megafossils.

TERTIARY SYSTEM

Buck Hill Volcanic Series

Principal features.—The Buck Hill volcanic series is chiefly tuff with subordinate conglomerate, sandstone, breccia, and fresh-water limestone and intercalated lava flows ranging from basalt to rhyolite (Fig. 3). The volcanic series lies with angular unconformity on the Boquillas formation in the Buck Hill quadrangle, and the base of the Pruett formation is marked by a limestone pebble conglomerate and conglomeratic arkosic sandstone. Intercalated with the Pruett beds are the Crossen trachyte, the Sheep Canyon basalt, and the Potato Hill andesite. These flows came into the Buck Hill region from the north, and their mapped boundaries probably represent their original southern limits fairly closely. The flows were eroded in part and weathered before being covered with ash. The Pruett formation is 900–1000 feet thick. In the northern part of the quadrangle the intercalated lavas make up 300–400 feet of the total thickness, but south of Crossen Mesa the whole thickness is tuff and related sediments.

The top of the Pruett formation is placed at the base of the Cottonwood Spring basalt flows which are variable in number and thickness, ranging from 50 to 325 feet or more. The flows are overlain by the Duff formation which is over 1000 feet thick and is chiefly rhyolitic tuff with some breccia and local conglomerate beds. The top of the Duff tuff is the base of the rhyolite capping Mitchell Mesa. West and southwest of the Buck Hill quadrangle the Mitchell Mesa rhyolite is overlain by a third major unit of tuff, tuffaceous sand, and thick flows of andesite-basalt totaling about 1000 feet, so that the Tertiary volcanic succession in this part of Trans-Pecos Texas has a minimum thickness of 3000 feet and may be as much as 4000 feet thick.

Age and correlation.—The Buck Hill volcanic series is of lower Tertiary age and is equivalent, in part at least, to volcanic successions in other parts of Trans-Pecos Texas. The Pruett formation is assigned tentatively to the Eocene; the Duff, to the Oligocene.

A collection of silicified gastropods from tuffaceous fresh-water limestone of the Pruett formation in Sheep Canyon in the Alpine quadrangle (Fossil locality, Pl. 6) was examined by Mr. Paul Bartsch of the U. S. National Museum for Hardin (1942). A *Physa*, the tips of *Goniobasis*, and the youthful form of *Cochliopa* were identified. Bartsch concluded that the *Goniobasis* specimens may belong to the

group including *tenera carterii*. *Goniobasis tenera carterii* Conrad is described from the lower Eocene of western Wyoming, western Colorado, and northern New Mexico (Henderson, 1935, p. 227). A few ostracods and charophytes in insoluble residues of the limestone were examined by Mr. R. E. Peck (Hardin, 1942) who found the internal molds of *Metacypris*, *Cypris*, and varieties of that genus, *Paracypris* and *Bairdiocypris*, etc.; and unidentifiable charophytes. None of these forms is diagnostic. Fresh-water limestone beds between flows of the Sheep Canyon basalt on the southern slope of Elephant Mountain contain a similar fauna. The assignment of the Pruett to the Eocene is on the basis of abundant specimens of *Goniobasis* in fresh-water limestone lentils in the upper part of the formation.

Low-spired fossil snails were collected from the lower Pruett tuff beds near the base of Crossen Mesa about 1 mile east of the Pruett ranch house and from near the base of Elephant Mountain. Similar large low-coiled snails, 2–3 inches in diameter, were collected by Graham (1942) from the Duff tuff at Church Mountain in the southeastern part of the Jordan Gap quadrangle. Land snails from the 02 ranch examined by Junius Henderson (Adkins, 1932, p. 514) are identical with the snail described by Cockerell (1914) as *Helix hesperarche*. Cockerell's specimen came from the Cope collection, and, although it was labelled "either Puerco or Torrejon, probably Torrejon", its original locality cannot be established. The specimen may have come from the Buck Hill volcanic series of Texas rather than from the Eocene Torrejon of New Mexico (Henderson, 1935, p. 134).

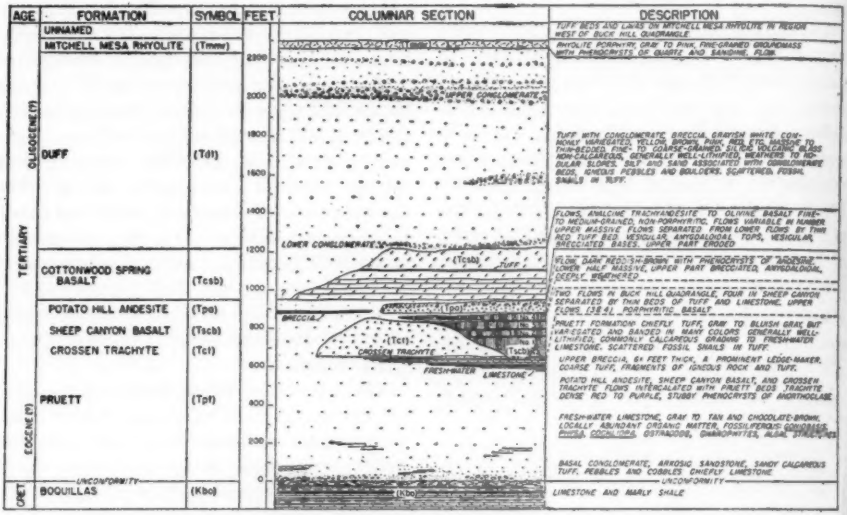
AGE
FORMATION
SYMBOL
FEET
COLUMNAR SECTION
DESCRIPTION
UNNAMED
MITCHELL MESA RHYOLITE (Tmrw)
DUFF (Tdf)
COTTONWOOD SPRING BASALT (Tcsb)
POTATO HILL ANDESITE (Tpa)
SHEEP CANYON BASALT (Tscb)
CROSSEN TRACHYTE (Tct)
PRUETT (Tpf)
UNCONFORMITY
BOQUILLAS (Kbq)
TERTIARY
OLIGOCENE(?)
EOCENE(?)
CRET.
2200
2000
1800
1600
1400
1200
1000
800
600
400
200
0
UPPER CONGLOMERATE
LOWER CONGLOMERATE
BRECCIA
TUFF
FRESH-WATER LIMESTONE
TUFF BEDS AND LAVAS ON MITCHELL MESA RHYOLITE IN REGION WEST OF BUCK HILL QUADRANGLE
RHYOLITE PORPHYRY, GRAY TO PINK, FINE-GRAINED GROUNDMASS WITH PHENOCRYSTS OF QUARTZ AND SANIDINE, FLOW
TUFF WITH CONGLOMERATE, BRECCIA, GRAYISH WHITE COMMONLY VARIEGATED, YELLOW, BROWN, PINK, RED, ETC., MASSIVE TO THIN-BEDDED, FINE TO COARSE-GRAINED, SILICIC, INCLUDING BEDS NON-CALCAREOUS, GENERALLY WELL-LITHIFIED, WEATHERS TO NODULAR BLOCKS, SILT AND SAND ASSOCIATED WITH CONGLOMERATE BEDS, IGNEOUS PEBBLES AND BOULDERS, SCATTERED FOSSIL SNAILS IN TUFF
FLOWS, ANALCIME TRACHYANDESITE TO OLIVINE BASALT FINE- TO MEDIUM-GRAINED, NON-PORPHYRITIC, FLOWS VARIABLE IN NUMBER UPPER MASSIVE FLOWS SEPARATED FROM LOWER FLOWS BY THIN RED TUFF BED, VESICULAR, AMYGDALOIDAL TOPS, VESICULAR, BRECCIATED BASES, UPPER PART ERODED
FLOW, DARK RED TO BROWN, WITH PHENOCRYSTS OF AUGITE, LOWER HALF MASSIVE, UPPER PART BRECCIATED, AMYGDALOIDAL, DEEPLY ERODED
TWO FLOWS IN BUCK HILL QUADRANGLE, FOUR IN SHEEP CANYON SEPARATED BY THIN BEDS OF TUFF AND LIMESTONE, FLOWS 138 FT., PORPHYRITIC BASALT
PRUETT FORMATION CHIEFLY TUFF, GRAY TO BLUISH GRAY BUT VARIEGATED AND BANDED IN MANY COLORS, GENERALLY WELL-LITHIFIED, COMMONLY CALCAREOUS GRADING TO FRESH-WATER LIMESTONE, SCATTERED FOSSIL SNAILS IN TUFF
UPPER BRECCIA, 64 FEET THICK, A PROMINENT LEDGE-MAKER, COARSE TUFF, FRAGMENTS OF IGNEOUS ROCK AND TUFF
POTATO HILL ANDESITE, SHEEP CANYON BASALT, AND CROSSEN TRACHYTE FLOWS INTERCALATED WITH PRUETT BEDS, TRACHYTE DENSE RED TO PURPLE, STUBBY PHENOCRYSTS OF ANORTHOCLASE
FRESH-WATER LIMESTONE, GRAY TO TAN AND CHOCOLATE-BROWN, LOCALLY ABUNDANT ORGANIC MATTER, FOSSILIFEROUS CONGLOMERATE, PISOLITE, OOLITE, OSTRACODS, CHAROPHYTES, ALGAL STRUCTURES
BASAL CONGLOMERATE, ARKOSIC SANDSTONE, SANDY CALCAREOUS TUFF, PEBBLES AND COBBLES CHIEFLY LIMESTONE
UNCONFORMITY
LIMESTONE AND MARLY SHALE

FIGURE 3.—COLUMNAR SECTION OF THE BUCK HILL VOLCANIC SERIES

The Duff formation is assigned to the Oligocene on the basis of a smaller gastropod associated with "*Helix*" *hesperarche* at Church Mountain and also found in other places in the Jordan Gap quadrangle. It is similar to a species occurring in the Oligocene Vieja volcanic series in the Tierra Vieja Mountains, western Presidio County (Fig. 2). The volcanic succession of the Tierra Vieja Mountains, Rim Rock country, was first studied by Vaughan (1900), who assigned it to the Cretaceous. In 1946 Mr. Bryan Patterson of the Chicago Natural History Museum collected vertebrates and gastropods from the lower Vieja tuff beds northwest of Porvenir (Fig. 2). Patterson (personal communication, 1947) states that the tuff beds are earliest Oligocene

with a fauna equivalent to that of the Chadron formation. The gastropods associated with the vertebrate fossils were identified as *Mesodon ("Helix") leidyi* (Hall and Meek) by Dr. L. S. Russell (personal communication, 1947) who states that the Duff tuff specimens probably are the same species.

A brief reconnaissance of the Porvenir area was made by Goldich with Patterson in November 1946. The basal conglomerate of the Vieja series lies with marked angular unconformity on the truncated steeply dipping Upper Cretaceous (Navarro) sediments. The volcanics have been deformed into broad folds with dips of 10°–15° on the limbs. There is a clear-cut angular unconformity between the Vieja tuff beds and the over-lying "lake beds" of Pliocene (?) age. Normal faults of considerable magnitude displacing the Vieja series and older rocks have been described by Baker (1941). The basal conglomerate contains many igneous pebbles and boulders, some 3 feet or more in diameter, in addition to limestone pebbles and cobbles. The conglomerate is 20–30 feet thick and is overlain by an equal thickness of tuff which is succeeded by thin flows of perlite and of trachyte. Only 3–4 feet of the perlite immediately below the trachyte is exposed. The trachyte, averaging 25–30 feet thick, is overlain by 400–500 feet of tuff capped by a massive igneous rock which Lord (1900) described as quartz pantellerite. Patterson's collections were made from the tuff beds above the trachyte.

South of the Tierra Vieja Mountains in the Chinati range (Fig. 2) Tertiary extrusives lie on Cretaceous and Permian sedimentary formations (Ross, 1943). In the Quitman Mountains in Hudspeth County a succession of flows, breccia, and tuff, 3500 feet thick, rests on folded and eroded Eagle Ford and older Cretaceous rocks. Huffington (1943) named this succession the Square Peak volcanic series and assigned it to the lower Tertiary, Eocene, or Oligocene.

In the Barrilla Mountains, Jeff Davis and Reeves counties, 60 miles north of the Buck Hill quadrangle, Tertiary volcanics rest unconformably on "clays of Pierre Cretaceous (Upson clay, Taylor marls) age" (Baker and Bowman, 1917, p. 119). Fossil leaves collected by Baker from the basal rhyolitic tuffs were correlated with the flora of the Raton and Denver formations (lower Eocene) by Berry (1919). From tuff beds within 200 feet of the base of the volcanic series on the Casey ranch 11 miles east of Balmorhea (Fig. 2) in the northeastern part of the Davis Mountains, Baker (1934, p. 151) collected a fossil tooth which R. A. Stirton identified as that of *Hyracodon*, a lower or middle Oligocene rhinoceras.

The Buck Hill volcanic series probably is equivalent in part to the Chisos beds (Udden, 1907b) which include about 2000 feet of thin-bedded tuffaceous sediments above the Tornillo clay in the Chisos Mountains. Udden concluded that there is no break between the Chisos beds and the Tornillo which contains a large quantity of weathered volcanic material (bentonite), and he assigned the Chisos beds to the Cretaceous. Adkins (1932, p. 508–514) considers the Tornillo as probably uppermost Cretaceous. The Chisos beds are now generally considered Tertiary.

Pruett Formation

Definition and description.—The Pruett formation, named for the Pruett ranch in the north-central part of the quadrangle, is principally volcanic tuff but includes conglomerate, tuffaceous sandstone and breccia, and tuffaceous fresh-water limestone. Interfingering with the tuff are flows of trachyte, basalt, and andesite.

The Pruett beds are widely distributed in the Buck Hill quadrangle and adjacent areas. East of Calamity Creek in Elephant Mountain the tuff is 500–600 feet thick. The volcanic succession in the Elephant Mountain area is described in a later section. South of the mountain in the eastern part of the quadrangle there are small remnants of the basal Pruett beds on the Boquillas. A thickness of 500 feet of tuff was measured (Appendix, Section No. 4) in the first canyon north of the southeastern end of Crossen Mesa 2 miles southwest of the Kokernot ranch house. The base of this section is covered by alluvium and may be 100–200 feet above the Cretaceous basement, because there is a minimum thickness of 650 feet of the Pruett in the southern scarp of the mesa. On the mesa additional tuff beds with intercalated lavas rest on the trachyte, and this

succession to the base of the Cottonwood Spring basalt is about 100 feet thick. If a thickness of 150 feet is assumed for the trachyte, a total thickness of at least 900 feet is indicated for the Pruett formation in the northern part of the quadrangle.

South of Crossen Mesa along the 02 ranch road the basal conglomeratic sandstone of the Pruett rests on Boquillas limestone. West of Whirlwind Spring the upper beds of the formation capped by the Cottonwood Spring basalt are exposed continuously, extending beyond the western boundary of the quadrangle into the Jordan Gap quadrangle. A 125-foot thickness of tuff was measured 1 mile south of Cottonwood Spring (Appendix, Section No. 5). Alluvium covers most of Green Valley, but the Pruett formation is exposed along Terlingua Creek, and along most of The Ditch and Calamity Creek Wash. The basal conglomerate and lower Pruett beds rest on the Boquillas limestone at Buck Hill and in a small graben 1 mile to the west.

The Pruett tuff is grayish white, bluish gray, greenish gray, buff, brown, pink, and red. Fragments commonly differ in color from the matrix, and speckled color combinations result. Locally hard, dense, and silicified layers grade to loose, friable ash, but on the whole the tuff is well indurated. Much of the Pruett is massive and thick-bedded, but thin-bedded, well-stratified material is common. Stratified, varicolored beds are strikingly exposed at Boat Mountain (Pl. 3, fig. 2), a spurlike prominence protected from erosion by a thick bed of breccia-conglomerate. Except where protected by the breccia-conglomerate or by the overlying flows of the Cottonwood basalt, the tuff is easily eroded and forms low rounded hills. It weathers to nodular forms, and slopes generally are littered with small hard pellets and nodules which roll easily making walking difficult.

There are few easily recognizable units in the Pruett formation which can be used for correlating separated sections, and faults are difficult to recognize and follow in the tuff unless one side of the fault is marked by an igneous rock or by Cretaceous strata. There are, however, three distinctive lithologic facies. They are tentatively proposed as members: (1) conglomerate and sandstone which are widespread at the base of the formation, (2) fresh-water limestone which in Crossen Mesa underlies the trachyte caprock, 500–600 feet above the base of the formation, and (3) breccia-conglomerate which occurs 60 feet below the top of the Pruett formation.

Basal conglomerate and sandstone.—The conglomerate at the base of the Pruett formation consists chiefly of well-rounded pebbles and cobbles of dense, light-tan to dark grayish-brown limestone with some marble, quartzite, petrified wood, and variously colored chert pebbles. It is characterized by flat pebbles of fibrous calcite which are pieces of *Inoceramus* shells. Pebbles of igneous rock are rare, requiring careful search, in contrast with their abundance in the breccia-conglomerate member near the top of the formation. The conglomerate grades up to coarse, cross-bedded conglomeratic sandstone.

The conglomerate, tuffaceous sandstone, and related beds are well exposed in a number of places in the eastern part of the quadrangle. Along the 02 ranch road south of Crossen Mesa where the conglomerate rests on thin-bedded Boquillas limestone, angular slabby blocks of the limestone are contained in the conglomerate which ranges from 2 inches to 1 foot in thickness and grades up to coarse-grained tuffaceous grit and sandstone. Approximately 2½ miles to the northwest in a small hill just below Crossen Mesa and three-fourths of a mile east of the Pruett ranch house, the lower tuff beds include a sandstone layer. The section measured in the small hill follows:

Unit	*Description*	*Thickness in feet*
6.	Tuff, hard, purplish with small white grains	30
5.	Tuff, hard, dense, pink to purplish	1
4.	Sandstone, tuffaceous, gray with a few black grains; well indurated (Sample No. 872). Pebbly layer 1–2 inches thick at base	7
3.	Tuff, gray, similar to unit No. 1	4
2.	Tuff, very hard, purplish with white rock and crystal fragments	2
1.	Tuff, gray to white; weathers to a hard nodular gravel. Base covered by alluvium	6
	Total thickness measured	50

PANORAMA OF NORTHERN BUCK HILL QUADRANGLE

Volcanic succession over 2000 feet thick is shown. Pruett tuff capped by Cottonwood Spring basalt forms lower mesas; Duff tuff capped by Mitchell Mesa rhyolite forms upper mesas. (Photograph by C. L. Baker)

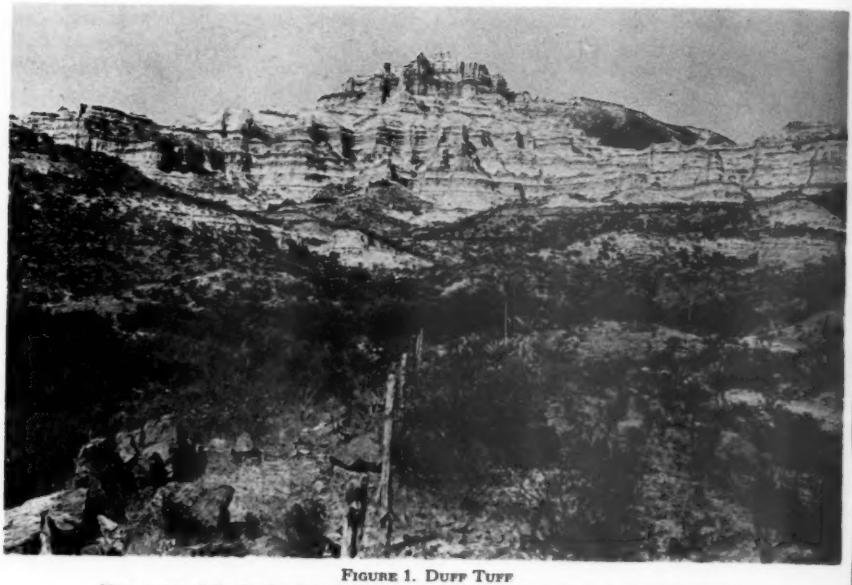

FIGURE 1. DUFF TUFF
Upper part of the Duff tuff in Jordan Gap quadrangle, 6 miles west of Mitchell Mesa. Clastic dike in left foreground.

FIGURE 2. PRUETT TUFF
Boat Mountain showing 200 feet of Pruett tuff capped by the upper breccia-conglomerate member.

PRUETT AND DUFF FORMATIONS

The basal conglomerate is well exposed in Hill 3927, 2½ miles northeast of Butcherknife Hill, where it is 6 feet thick and grades up to tuffaceous pebbly sandstone, 15–20 feet thick. Grains of biotite and other dark minerals and fragments of dark chert give the sandstone a speckled appearance. Good exposures of the basal Pruett beds also are found in the vicinity of Butcherknife Hill. On the northwest side of the hill the conglomerate, sandstone, and breccia dip 25° NW. On the east side, the lower part of the section is covered, and the conglomerate is not exposed, but a 10-foot bed of sandstone is included in the calcareous tuff beds. The section follows:

Unit	Description	Thickness in feet
5.	Interval covered with float of dense basalt exposed at top of Butcherknife Hill. In part, Butcherknife basalt..	77
	Pruett tuff; basal conglomerate member:	
4.	Tuff, containing thin limestone beds and nodules near top grading down to calcareous tuff and sandy tuff at base of interval.....	17
3.	Sandstone, light gray, very coarse-grained, friable, tuffaceous, and calcareous. (Sample No. 1231).....	10.5
2.	Tuff, gray, sandy.....	5.5
1.	Tuff, red, fine-grained, base covered by float and alluvium.....	2
	Total thickness measured.....	112

At Buck Hill the syenite which now caps the hill intruded the lower Pruett beds. The following section was measured on the southeastern side of the hill:

Unit	Description	Thickness in feet
6.	Syenite to top of hill	
	Pruett formation; basal conglomerate member:	
5.	Tuff, hard red.....	5
4.	Sandstone, calcareous, with layers of tuffaceous limestone.....	10
3.	Conglomerate, limestone pebbles in a calcareous sandstone matrix.....	3
	Boquillas limestone:	
2.	Limestone, thin-bedded, fine-grained.	2
1.	Limestone, thin- to thick-bedded, dense, black, with poorly preserved pelecypods. Base covered.....	8
	Total thickness measured.....	28

In a large gully south of Bench Mark 4203 on Buck Hill, a short distance west of the locality of the preceding section, 30 feet of the lower Pruett is exposed. The section follows:

Unit	Description	Thickness in feet
5.	Syenite to top of hill.....	50
	Pruett formation; basal conglomerate member:	
4.	Tuff, red and yellow, fine-grained....	15
3.	Sandstone, tuffaceous, calcareous....	10
2.	Conglomerate and conglomeratic sandstone.....	5
	Boquillas limestone:	
1.	Limestone, massive beds at top with thin, slabby limestone below. Covered at base for 82 feet to Cretaceous limestone flat south of Buck Hill. Thickness including covered part....	126
	Total thickness measured.....	206

In the Buck Hill quadrangle the basal conglomerate and sandstone generally do not exceed 15 feet in thickness. The overlying tuff beds are sandy and calcareous and commonly include one or more sandstone layers. Field relations indicate that the beds of sandstone in the tuff at Butcherknife hill and in the small hill south of Crossen Mesa are not many feet above the base of the formation. Accordingly they are included in the basal conglomerate member which is defined to include the conglomerate, sandstone and arkose, and sandy and calcareous tuff in the lower 40–50 feet of the Pruett formation.

Fresh-water limestone.—Massive beds of fresh-water limestone occur in the Pruett formation just below the Crossen trachyte. In Sheep Canyon in the Alpine quadrangle (Pl. 6) these tuffaceous limestone and highly calcareous tuff beds are 300 feet thick in the spur between Sheep Creek and Calamity Creek. The limestone ranges from grayish white to tan and chocolate brown. Much of it is in beds up to 10 feet thick, but locally beds a few inches thick are interbedded with calcareous tuff or are separated by shaly partings. The limestone is fossiliferous, and the silicified gastropods stand out on weathered surfaces. Algal structures are common.

In Sheep Canyon the limestone is overlain by the Crossen trachyte and by the Sheep Canyon basalt where the trachyte has been eroded. In the eastern scarp of the mesa east

of Calamity Creek and north of Elephant Mountain, the fresh-water limestone lies both below and above the Crossen trachyte where the latter has been thinned, presumably by erosion. Beds of the tuffaceous limestone, 40 feet thick, are overlain by the Sheep Canyon basalt, and the three lower basalt flows are separated by beds of the fresh-water limestone (Appendix, Section No. 6). The limestone is in part older than the trachyte, and in part its deposition spans the interval of extrusion of the Crossen trachyte and the Sheep Canyon basalt lavas.

Upper breccia-conglomerate.—Tuff breccia and conglomerate in the upper part of the Pruett formation, approximately 60 feet below the Cottonwood Spring basalt, are designated the upper breccia-conglomerate member. It may be a single thick, well-cemented, and coherent layer or it may consist of several layers about 1 foot thick separated by thin beds of coarse tuff. The breccia contains angular fragments of a variety of igneous rocks with some pieces of chert and jasper. Generally the pieces are less than half an inch in greatest dimension and exhibit structures and textures such as spherulitic, eutaxitic, fine-grained dense or vesicular, and porphyritic with small crystals of sanidine and quartz. Some dark-red vesicular fragments suggest basic igneous rocks, but most are silicic. Locally some of the pieces are rounded and attain a foot or more in diameter.

The breccia-conglomerate is cemented with calcareous tuffaceous material and is resistant to erosion forming a bench in the Pruett formation below the basalt-capped hills and mesas west of Whirlwind Spring and north of the 02 ranch road. At Whirlwind Spring the bed, 2–6 feet thick, is composed of angular igneous fragments which generally are less than half an inch across. The layer is much thicker at Boat Mountain where large slabby pieces have slumped off from the caprock ledge and cover the base of the slope. In this slumped material along the road there are more or less rounded igneous boulders as much as 3 feet in diameter. West of Boat Mountain and south of elevation 4054 feet (Appendix, Section No. 5) the breccia-conglomerate is 2–6 feet thick with a considerable range in size of fragments.

Good exposures of the breccia-conglomerate and tuff beds above it are found along Duff Creek in the western part of the quadrangle and can be traced westward into the Jordan Gap quadrangle where the breccia-conglomerate member is as much as 40 feet thick. In the Duff Creek area the member is 2–6 feet thick and is composed of angular to well-rounded pieces many of which are 2–3 inches in diameter. The proportion of fragments to tuff matrix is large. The following section was measured on the west slope of Hill 4325 east of Duff Creek:

Unit	*Description*	*Thickness in feet*
Cottonwood Spring Basalt:		
8.	Basalt, fine-grained, massive, dark-green	
Pruett formation:		
7.	Tuff, dense, hard, baked, dark-red	3
6.	Ash, soft, earthy, weathered (?), with no apparent stratification, red	30
5.	Tuff, thick-bedded, red with white specks	11
4.	Tuff, coarse, poorly bedded but in thick rough layers, mottled gray and pink	7
3.	Tuff, thin-bedded, pink and gray	11
2.	Upper breccia-conglomerate member, resistant bed of large volcanic fragments in a five-grained red tuff matrix	2
1.	Tuff to base of slope	
	Total thickness measured	64

The breccia-conglomerate member was originally extensive, covering many square miles. It caps the Pruett in the faulted block southwest of the Pruett ranch house. A thin bed of breccia with small angular gray to reddish-brown fragments in a porous light-grayish-pink to pink matrix caps 50 feet of tuff on the trachyte in Hill 5202 on Crossen Mesa. This breccia occupies a position below the Cottonwood Spring basalt which stratigraphically is analogous to that of the breccia-conglomerate in the downfaulted area southwest of the mesa and is considered a remnant of the breccia-conglomerate member. Northwest of Hill 5202 on Crossen Mesa the breccia-conglomerate was not found, and its position in the succession is occupied by the Potato Hill andesite flow (Fig. 3).

Composition.—The greater part of the Pruett

formation is fine-grained tuff, commonly calcareous, and locally silicified. The lower tuff beds above the basal conglomerate-sandstone member generally are calcareous and commonly contain thin layers and nodules of limestone. Part of the calcium carbonate probably was derived from the erosion of the exposed Cretaceous formations, but a large part probably was precipitated in the lakes and ponded water in which the volcanic ash was deposited. In the northern part of the Buck Hill quadrangle and in the Alpine quadrangle thick beds of fresh-water limestone in the upper part of the formation were deposited in ponds. Abundant algal structures, charophytes, and fossil gastropods suggest that organic processes were important in the accumulation of the limestone. The limestone is tuffaceous and grades to calcareous tuff.

The tuff beds in the uppermost part of the formation generally contain less calcium carbonate than the lower beds. Thin sections show no recognizable glass, but relict structures suggest devitrified shards. Fragments of quartz and feldspar are variable in abundance. A light-gray to cream-colored sample from poorly stratified tuff below the Potato Hill andesite at Hill 4975 on Crossen Mesa was analyzed (Table 2, No. 1). The tuff, 37 feet thick at this locality, is overlain by 40 feet of andesite succeeded by 26 feet of tuff to the base of the Cottonwood Spring basalt (Appendix, Section No. 7), so that the analyzed sample was taken 70 feet below the top of the Pruett formation. The sample is similar to an analyzed sample from the younger Duff tuff of Mitchell Mesa (Table 2, No. 2). Both samples are considerably altered as indicated in the large contents of water. Despite the greater alteration suggested by its larger water content, the Pruett tuff sample has a larger relative amount of lime which in the absence of carbon dioxide suggests that the parent material of the Pruett tuff represented by the analyzed sample probably was more basic than that of the Duff tuff.

Part of the analyzed sample was crushed, and the heavy minerals separated with bromoform (D = 2.86). The heavy-mineral fraction is small and consists principally of magnetite partially altered to hematite and limonite. Chlorite and sphene are the most abundant nonopaque minerals. The chlorite is in small rounded flakes; the sphene commonly shows crystal faces. A few grains of apatite, aegirine-augite, rutile, biotite, muscovite, and zircon were noted. A sample of gray calcareous tuff from below the trachyte in the ravine south of Hill 5202 contains similar heavy minerals, but fresh biotite, zircon, and apatite are relatively more abundant. Very hard, dense, purple, silicified tuff in thin layers about halfway up the southern scarp of Crossen Mesa is composed of shards of silicic volcanic glass with minor quartz and feldspar, a few rock fragments, and a little calcite.

TABLE 2.—*Chemical analyses of samples of tuff from the Pruett and Duff formations*
(S. S. Goldich, analyst)

	1	2
SiO_2	69.73	70.57
Al_2O_3	12.06	12.19
Fe_2O_3	1.06	1.29
FeO	.13	.09
MgO	.88	.60
CaO	2.16	1.43
Na_2O	.78	1.80
K_2O	4.17	5.09
H_2O+	5.05	3.36
H_2O-	3.31	2.84
CO_2	.01	.14
TiO_2	.18	.22
P_2O_5	.04	.08
MnO	.01	.01
BaO	.13	.04
SrO	.09	n.d.
S	.00	.00
	99.79	99.75
Sp. gr.	2.368	2.421

1. Tuff from the Pruett formation below Potato Hill andesite at Hill 4975 on Crossen Mesa. Sp. gr. 24°/4°.
2. Tuff from the Duff formation, approximately 100 feet below Mitchell Mesa rhyolite, Mitchell Mesa. Sp. gr., 26.6°/4°.

The coarser clastic materials of the Pruett formation include conglomerate, breccia, and arkosic sandstone. The basal conglomerate consists chiefly of well-rounded pebbles and cobbles of limestone with more angular pebbles

of chert, jasper, and petrified wood. Pebbles of marble and quartzite are common, but igneous rock pieces are rare. The related sandstone contains abundant orthoclase and

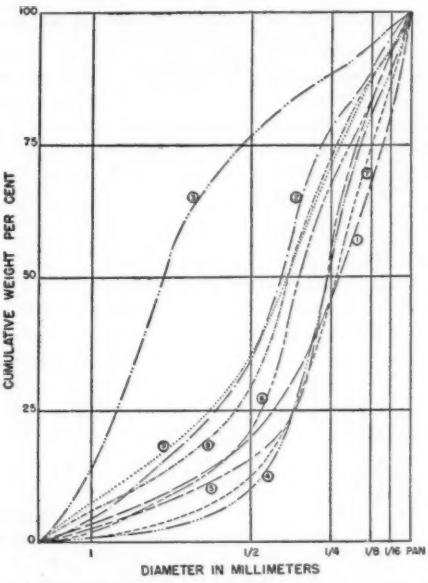

FIGURE 4.—CUMULATIVE CURVES REPRESENTING TEXTURAL ANALYSES OF SANDSTONE SAMPLES FROM THE PRUETT FORMATION

silicic plagioclase. In addition to the lower beds there are lentils and thin layers of conglomerate and sandstone at higher levels. On the northwest slope of Hill 4140 south of the 02 ranch house a bed of conglomerate 6½ feet thick is overlain by 5 feet of sandstone. The conglomerate is composed of rounded limestone and tuff pebbles. These beds occur in the tuff 70 feet above the base of the hill and 30 feet below the base of the overlying basalt flows which are assigned to the Cottonwood Spring basalt, so that the sandstone and conglomerate are near the top of the Pruett formation. The beds appear to be local and could not be traced in other parts of the hill. Three miles south of Hill 4140 in the bed of Terlingua Creek tuffaceous sandstone is exposed along an east-west fault which is the western extension of the Chalk Draw fault.

Nine samples of sandstone from the Pruett formation from scattered localities in the quadrangle were treated with dilute hydrochloric acid to dissolve the calcareous cement. The loss in weight in the acid ranged from 26 to 61 per cent. The clay and silt fraction ranges from 1.6 to 4.2 per cent, and the sand fraction from 37 to 71 per cent. Constants for the sand fractions of the samples from cumulative curves (Fig. 4) are given in Table 3. These include the median diameter (Md) and coefficients of sorting (So) and of skewness (Sk). The samples are calcareous sandstones in which the medium to coarse sand is characterized by a high degree of sorting. There is no appreciable difference between sample No. 8 from the upper part of the formation and the samples from the basal conglomerate member. Sample No. 5 from the southwest slope of Butcherknife Hill is an arkose containing abundant silicic feldspar, quartz, and rock fragments cemented by clear granular calcite. Some dense, darker calcite with sections of Foraminifera and structures suggesting fossil shells appears to be pieces of fragments derived from the Cretaceous limestones. Small fragments of ophitic and trachytic rocks, chert, and chalcedony are minor, and the principal constituents in addition to the calcite and limestone fragments are abundant angular orthoclase, silicic plagioclase, and quartz. Accessory minerals are biotite, muscovite, tourmaline, and a radiating fibrous mineral which may be a zeolite.

Thin sections of the upper breccia-conglomerate from localities south of Cottonwood Spring and in Duff Creek show fragments of spherulitic, eutaxitic, dense, and porphyritic varieties of rhyolite, and trachytic and ophitic volcanics. The ophite consists of minute plagioclase laths in an indeterminable reddish-brown matrix. The fragments are cemented with a selvage of fine-grained material containing grains of quartz and alkalic feldspar.

Stratigraphic relations.—The Pruett formation rests unconformably on the Boquillas formation and is overlain by the Cottonwood Spring basalt. Deposition of the tuff and related sediments was locally interrupted by invasions of lavas which moved into the area from the north. The Crossen trachyte, which rests on tuff and fresh-water limestone of the

upper Pruett, was eroded and removed from large areas, so that the younger Sheep Canyon basalt flows came to rest on the fresh-water limestone member and overrode the erosion-

TABLE 3.—*Bulk composition and textural constants for samples of sandstone from the Pruett formation*

	Weight per cent								
	1	2	3	4	5	6	7	8	9
Bulk composition:									
Loss in dil. HCl	33.2	59.4	27.8	41.6	40.4	60.8	30.2	26.2	36.0
Clay and silt	1.6	3.4	1.6	4.2	4.0	2.6	6.0	3.0	2.0
Sand	65.2	37.2	70.6	54.2	55.6	36.6	63.8	70.8	62.0
Constants:									
Md (mm.)	0.23	0.41	0.78	0.26	0.26	0.35	0.23	0.38	0.39
So	2.17	1.49	1.32	1.52	1.47	1.56	1.71	1.54	1.70
SK	.79	1.01	.81	.87	.88	.73	.85	.87	.91

1. Above conglomerate in small hill below Crossen Mesa, 1 mile east of Pruett ranch house. Cat. No. 872.
2. Above basal conglomerate, along O2 ranch road, 1¼ miles northeast of Whirlwind Spring. Cat. No. 869.
3. From basal conglomerate member, east side of Butcherknife Hill. Cat. No. 1231.
4. Same as 3, west side of Butcherknife Hill. Cat. No. 830-A.
5. Same as 3, southwest slope of Butcherknife Hill. Cat. No. 831.
6. Above basal conglomerate, along road to Sid Place, southwest side of fault, approximately 2 miles west of Butcherknife Hill. Cat. No. 1232.
7. From bed of Calamity Creek Wash, half a mile northwest of sample No. 6. Cat. No. 1230.
8. Above conglomerate bed, approximately 70 feet above base on northwest slope of hill half a mile south of O2 ranch house. Cat. No. 883.
9. From bed of Terlingua Creek, along fault west of Straddlebug Mountain. Cat. No. 1244.

thinned trachyte (Fig. 3). A third invasion of porphyritic andesite followed. In the northern part of the quadrangle the basal Cottonwood Spring flow rests on deeply weathered Potato Hill andesite or on tuff beds on the andesite, but south of the Walnut Draw fault the basal flow rests on soft, earthy, red ash which probably represents deeply weathered Pruett tuff. Breaks in the deposition of the Pruett, other than those caused by the lavas, are difficult to determine. The upper breccia-conglomerate, 60 feet below the top of the formation, marks a break, the magnitude of which is not known, but in this member is the first appearance of large igneous cobbles and boulders.

Crossen Trachyte

Definition and description.—The massive flow of porphyritic trachyte capping Crossen Mesa in the north-central part of the Buck Hill quadrangle is designated the Crossen trachyte. It covers approximately 20 square miles on the mesa (Pl. 1) and also caps extensive areas to the north in the Alpine quadrangle (Pl. 6). The flow came into the Buck Hill region from the north or northwest. No trace of the trachyte is found in Elephant Mountain east of Crossen Mesa, but it may have been removed by erosion. Northwest of Elephant Mountain (Pl. 6) the trachyte was eroded from a large area by an ancient (Eocene?) stream.

The fresh trachyte is hard and brittle with an even to subconchoidal fracture. Abundant stubby phenocrysts of glassy feldspar are evenly distributed in a fine-grained, grayish to reddish-brown groundmass. Near the southern rim of Crossen Mesa the upper part of the flow is vesicular with many fillings of chalcedony. The vesicles are strongly elongated, generally northwest-southeast, accentuating the flow lines and the alignment of the phenocrysts. The vesicular top of the flow locally was broken and recemented to form a prominent flow breccia which is well exposed in the bed of Goat Creek on the mesa

and in Walnut Draw on the western edge of Crossen Mesa.

The present surface of the Crossen trachyte is essentially a dip slope, dipping 1°–2°, N. 20° W., from which erosion has removed much

TABLE 4.—*Chemical analysis of Crossen trachyte*

From the upper part of the flow, head of first gully north of the southeastern end of Crossen Mesa
(S. S. Goldich, analyst)

SiO_2	71.17
Al_2O_3	12.64
Fe_2O_3	4.28
FeO	.11
MgO	.13
CaO	1.05
Na_2O	3.91
K_2O	5.05
H_2O+	.23
H_2O-	.12
CO_2	.19
TiO_2	.56
P_2O_5	.07
MnO	.14
BaO	.08
SrO	n.d.
S	tr.
	99.73
Sp. gr. (24°/4°)	2.648

of the brecciated vesicular top. In the cliff on the southern edge of the mesa the trachyte is about 120 feet thick, and a thickness of 115 feet is estimated (Appendix, Section No. 4) in the first gully north of the southeastern tip of the mesa. The flow is considerably thicker to the northwest, averaging 150 feet.

The trachyte weathers to a rusty-brown pitted surface, the pits resulting from the weathering out of the phenocrysts. Some of the weathered rock is bleached to a light yellow or yellowish gray, and as this type is found in place beneath calcareous tuff it may represent ancient (Eocene?) weathering. The vertical jointing of the flow favors the development of vertical walls or cliffs on the edge of the mesa. Marginal fragmentation of the trachyte yields abundant debris that covers the more gentle slopes of the less coherent underlying tuff.

Petrography.—The fine-grained groundmass is not readily resolved even under high magnification. It consists principally of microlites of silicic feldspar with minor amounts of altered ferromagnesian mineral, which originally probably was aegirine-augite, and of magnetite and apatite. Secondary iron oxides and a chloritic mineral are the chief alteration products, and the red color of the rock is derived from the abundant finely disseminated hematite.

The feldspar phenocrysts are rounded and irregularly corroded. They are chiefly anorthoclase with delicately developed quadrille twinning and moderately large 2V. Clear untwinned feldspar with small (-) 2V is sanidine. The phenocrysts are much altered to carbonate, hematite, and a chloritic mineral. These secondary products in part were developed at the expense of original ferromagnesian minerals included in the feldspar phenocrysts. Carbonate pseudomorphs with outlines marked by films of hematite and with small veinlets of hematite suggest original olivine. A few larger grains of the original pyroxene largely altered to a dark-green mineral appear to be aegirine-augite. Grains of granular quartz probably are secondary, and chalcedony was introduced.

Chemical analysis.—The trachyte is characterized by an unusually high silica content of 71 per cent (Table 4), largely as a result of the introduction of secondary quartz and chalcedony. The relatively large content of ferric oxide reflects the oxidation of the iron-bearing minerals to hematite.

Stratigraphic relations.—The Crossen trachyte rests on tuff and fresh-water limestone of the Pruett formation. In the vicinity of Potato Hill it is overlain by tuff beds assigned to the upper part of the Pruett which are succeeded by the Potato Hill andesite. Near the northern boundary of the quadrangle a flow of the Sheep Canyon basalt rests directly on the trachyte, and farther north in the Alpine quadrangle (Hardin, 1942) the trachyte was eroded by an ancient stream, the channel of which subsequently was filled with flows of the Sheep Canyon basalt. Small hills or erosional remnants of the Crossen trachyte were surrounded by the flows, and these now appear as inliers in the Sheep Canyon basalt (Pl. 6). The stratigraphic relationships of the

Crossen trachyte, the Pruett beds, and the Sheep Canyon basalt flows are shown schematically in Figure 3.

Sheep Canyon Basalt

Definition and description.—Basalt flows intercalated with fresh-water limestone and tuff beds of the Pruett formation are referred to the Sheep Canyon basalt. These are younger than the Crossen trachyte and older than the Potato Hill andesite. The flows are named for Sheep Canyon, a re-entrant of Calamity Creek valley in the Alpine quadrangle approximately 3 miles north of the Buck Hill quadrangle (Pl. 6).

The Sheep Canyon flows originally were extensive in the northern part of the Buck Hill quadrangle but have been eroded from large areas. The basalt is now found on Crossen Mesa and in the downfaulted block southwest of the mesa. The southernmost outcrop of the flows is along an east-west fault approximately 1 mile south of the Pruett ranch house. The flows cover large areas in the Alpine quadrangle where they are much thicker. In Sheep Canyon at least four flows are separated by thin beds of tuff and tuffaceous limestone.

The basalt typically is dark greenish black, even-textured, fine- to medium-grained, locally containing large, glassy, green plagioclase phenocrysts, many of which are 1 inch long. On relatively steep slopes the basalt commonly develops a platy structure on weathering; on gentle slopes and in the bottoms of gullies spheroidal structures result. The vesicular tops of the flows contain abundant fillings of quartz, chalcedony, and calcite and commonly are a distinctive green. In many places small flats and benches are developed at the contact of two flows.

At the type locality in Sheep Canyon in the section measured by Hardin (Appendix, Section No. 6), the basalt flows are 234 feet thick. The upper basalt flow in this section is overlain directly by the Potato Hill andesite, but on Crossen Mesa in the Buck Hill quadrangle a thickness of 20 feet or more of tuff separates the Sheep Canyon basalt from the Potato Hill andesite. A section measured on a little hill north of Potato Hill and just south of the northern boundary of the quadrangle follows:

Unit	*Description*	*Thickness in feet*
Pruett formation:		
4.	*Potato Hill andesite intercalation*	
3.	Tuff, gray, well stratified, interbedded with purplish-red ash speckled with white fragments; upper ⅔ feet baked, dense, red tuff	27
2.	*Sheep Canyon basalt intercalations:*	
	b. Basalt flow, black fine-grained rock with vesicular top altered green; weathers to flattish biscuitlike forms. Red baked ash included in thin flow breccia at base of flow	34
	a. Basalt flow, uniform black, fairly coarse-grained but not porphyritic. Well-developed vesicular top	21
1.	*Crossen trachyte intercalation:* Red porphyritic trachyte breccia	
	Total thickness measured	82

Half a mile south of this hill the upper flow only is present. It is 27 feet thick, coarse-grained and porphyritic. The vesicular base of the flow rests on baked, dense, red tuff grading down to well-stratified gray tuff resting on trachyte breccia. The tuff is 13 feet thick. The green vesicular top of the basalt flow is overlain by stratified gray ash capping the low hill. To the west this ash is overlain by the Potato Hill andesite. East of Goat Creek and southwest of elevation 5016 feet, the Sheep Canyon basalt is intercalated with ash resting on the Crossen trachyte. At this point the basalt is 30 feet thick but feathers out to the southeast, and on the southern scarp below elevation 4975 feet no trace of the Sheep Canyon basalt is found.

In the spur west of Goat Creek two flows of the Sheep Canyon basalt rest on 11 feet of tuff and are overlain by 20 feet of tuff beneath the Potato Hill andesite. The flows are of equal thickness, totaling 65 feet, and have well-developed vesicular tops. A prominent flat or bench on the southeastern end of the spur roughly marks the top of the first flow.

South of Crossen Mesa the Sheep Canyon basalt is exposed in gullies in an area approximately 1 mile square. This area is bounded on the east and south by the Cottonwood Spring basalt which is downfaulted against

the Pruett tuff. At this locality a flow 30 feet thick lies in well-stratified, grayish-white tuff. The top is amygdaloidal with many fillings of chalcedony and calcite. The flow was

TABLE 5.—*Chemical analyses of Sheep Canyon basalt from Buck Hill and Alpine quadrangles*
(S. S. Goldich, analyst)

	1	2	3
SiO_2	44.33	45.65	45.97
Al_2O_3	14.69	16.48	17.26
Fe_2O_3	4.27	6.92	2.84
FeO	8.68	6.40	9.18
MgO	4.88	4.05	5.63
CaO	7.00	7.48	7.87
Na_2O	3.46	3.38	4.33
K_2O	2.15	1.42	1.00
H_2O+	2.15	2.10	1.57
H_2O-	1.09	1.39	.20
CO_2	.07	.37	.00
TiO_2	3.64	3.21	2.87
P_2O_5	2.99	.87	.96
MnO	.23	.18	.17
BaO	.17	.09	.07
SrO	.14	.09	.10
S	.00	.00	.00
	99.94	100.08	100.02
Sp. gr.	2.818	2.829	2.874

1. Basalt flow in gully, 1450 feet N. 48° W. of Hill 4321′, south of Crossen Mesa. Cat. No. 841. Sp. gr., 25.2°/4°.
2. Porphyritic basalt, upper of two flows on Crossen Mesa, 1 mile north of Potato Hill. Cat. No. 832. Sp. gr., 25.8°/4°.
3. Porphyritic basalt in gully, east side of Elephant Mountain, Alpine quadrangle; 100 feet below base of syenite. Cat. No. E. M. 14. Sp. gr., 22°/4° (See Pl. 6).

weathered following extrusion of the lava, and fragments of weathered vesicular basalt are contained in the overlying gray tuff which grades upward to pink and grayish-white stratified volcanic ash. Narrow (2-inch) veins of massive aragonite cut the basalt.

Petrography.—In Sheep Canyon the basalt flows are rather uniform both in texture and in composition. The groundmass is dominantly feldspathic and distinctly diabasic. The plagioclase is labradorite (An_{55}), and the large crystals commonly are zoned with cores as calcic as An_{60}. Zonal inclusions of magnetite and pyroxene in the feldspar phenocrysts are common. Augite, olivine, and magnetite are the other essential primary minerals, and apatite is the chief accessory mineral. Olivine occurs in about equal amounts in all the flows. Hardin (1942) estimates the composition of the olivine from measurements of the optic angle to be approximately Fo_{70} Fa_{30}. The olivine is considerably altered to antigorite which shows the polygonal outline of the olivine crystals, but there are some residual olivine grains. The amount of augite is variable. Magnetite is abundant—10–15 per cent of the rock. Analcime, calcite, sericite, fibrous zeolites, and clay minerals occur in varying amounts.

Thin sections of the basalt from the Buck Hill quadrangle show the rock to be generally similar to the flows in Sheep Canyon, but a few special features are worthy of mention. Samples of the flows north of Potato Hill show zoning in the plagioclase which is the reverse of the normal zoning described above, and rounded central zones are more sodic than the outer ones. The zones commonly are marked by lines of euhedral crystals of magnetite. A silicic feldspar, probably orthoclase, in the groundmass is interstitial and marginal to the plagioclase. The serpentinous alteration products of olivine include olive-green fibrous material and a reddish-brown flaky mineral, probably bowlingite. Calcite is an abundant secondary mineral. The pyroxene is distinctly purplish and pleochroic indicating a titaniferous augite. Samples of greenish-gray basalt collected southwest of Hill 5106 feet are fine- to medium-grained with a diabasic to granular texture. Feldspar makes up 70 per cent or more of the rock. Olivine and its alteration products do not exceed 5 per cent of the rock, and pyroxene, 10–15 per cent. The abundance of silicic feldspar is striking, and the rock is more nearly a latite than basalt.

The basalt in the downfaulted block south of Crossen Mesa resembles the flows north of Potato Hill. Analcime is not abundant. Apatite is exceedingly abundant in small prismatic crystals.

Chemical analysis.—Three analyses of samples of the Sheep Canyon basalt (Table 5) are similar, but in relative abundance of minor constituents, P_2O_5, MnO, BaO, and SrO, the

FIGURE 1. LARGE MESA CAPPED BY COTTONWOOD SPRING BASALT
Southwest of Cottonwood Spring. Flat-topped McKinney Mountain, Jordan Gap quadrangle, in distance.

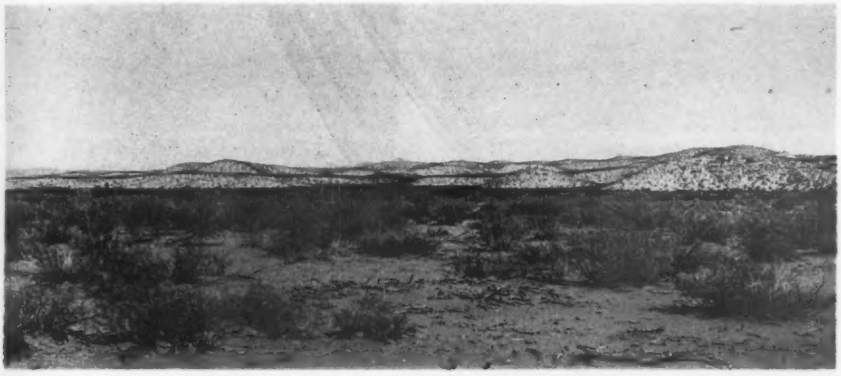

FIGURE 2. UPPER CRETACEOUS BOQUILLAS FORMATION
Rounded hills of the Boquillas formation south of Elephant Mountain.

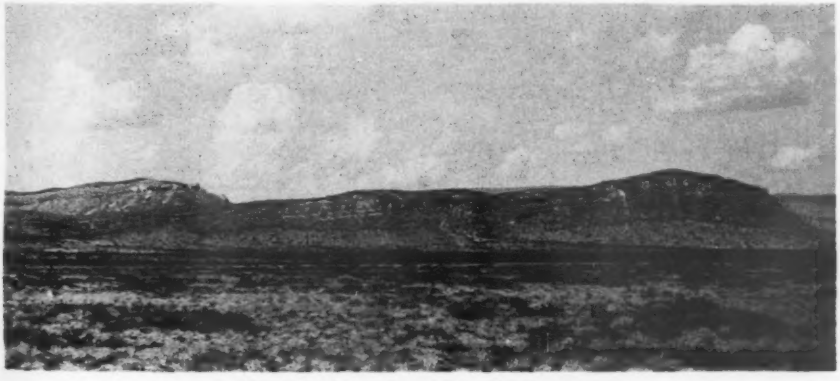

FIGURE 3. LOWER CRETACEOUS FORMATIONS
The lower scarp on the right is formed by the Georgetown limestone; upper scarp on the left, by the Grayson clay overlain by the Buda limestone. (Agua Fria quadrangle).

COTTONWOOD SPRING BASALT AND CRETACEOUS FORMATIONS

FIGURE 1. POTATO HILL ANDESITE
Deeply weathered andesite breccia in Walnut Draw; man is standing on contact of Cottonwood Spring basalt with tuff bed on the andesite.

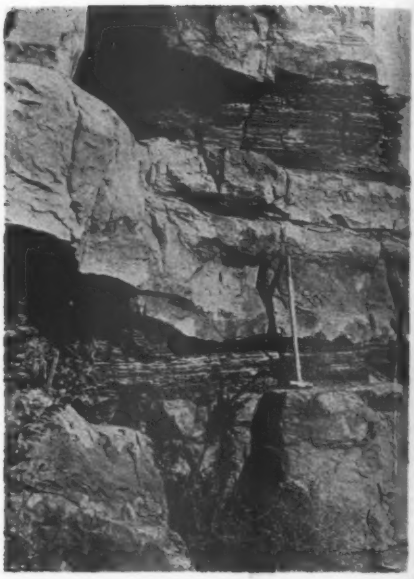

FIGURE 2. SYENITE AT BUCK HILL
Inclusions of thin-bedded Boquillas limestone in syenite.

FIGURE 3. SLICKENSIDES ON FAULT PLANE AT BUCK HILL
Syenite on right is downthrown against basal conglomerate member of the Pruett formation on left.

Elephant Mountain porphyritic basalt more closely resembles the porphyritic basalt sampled north of Potato Hill on Crossen Mesa. The large percentage of titania in the analyzed samples is noteworthy.

Potato Hill Andesite

Definition and description.—The Potato Hill andesite, named for Potato Hill on Crossen Mesa, is a dark reddish-brown porphyritic flow and flow breccia in the upper part of the Pruett tuff. The andesite caps a small hill 500 feet southeast of Potato Hill. Northwest of Potato Hill the flow makes a narrow ridge, and its outcrop continues to the northern boundary of the quadrangle on either side of the basalt-capped hills that crown Crossen Mesa. The andesite is exposed in the spur between Goat Creek and Walnut Draw and in the faulted blocks which make Walnut Draw graben on the western edge of Crossen Mesa, but apparently the flow did not extend beyond the southern limit of the Crossen trachyte, because it is not found in the downfaulted block south of the Walnut Draw fault zone.

The Potato Hill andesite is easily recognized by its color and prominent plagioclase phenocrysts. The tabular crystals, up to 1 inch in length, are fresh and glassy. They are contained in a fine-grained, reddish-brown groundmass or in a coarser granular grayish-brown matrix. Locally the andesite is altered to a greenish color. The lower part of the flow, although vesicular, is massive; the upper half is a flow breccia. Amygdules of quartz, chalcedony, and calcite are abundant. The upper brecciated part weathers more readily than the massive andesite, and small flats are developed at the change of structure in the flow. The flow was deeply weathered after extrusion and before the succeeding ash was deposited.

At Potato Hill the andesite is 31 feet thick. It lies on approximately 50 feet of tuff which rests on the Crossen trachyte. The andesite is overlain by 25 feet of hard, dense, red tuff. Below elevation 4806 feet, 6000 feet north of Potato Hill, the andesite is only 18 feet thick. It is overlain by 29 feet of well-stratified tuff below the Cottonwood Spring basalt. The relative thinness of the andesite at this point may be the result of weathering rather than original thinning of the flow because in the Alpine quadrangle the andesite is up to 70 feet thick with appreciable flow breccia. In a measured section (Appendix, Section No. 7) in Hill 4975, 1½ miles northwest of Potato Hill the andesite is 40 feet thick, and this thickness holds fairly constant to the northwest. In the spur on Crossen Mesa between Goat Creek and Walnut Draw a small bench or flat marks the break between the massive and brecciated phases of the andesite. The section measured on the southeastern end of the spur follows:

Unit	*Description*	*Thickness in feet*
	Cottonwood Spring basalt:	
7.	Basalt flow; fine-grained, jointed; weathers to reddish-brown slabs. To top of hill	44
	Pruett formation	
6.	Tuff; red, fine- to medium-grained, grading down to weathered andesite, contact approximate	10
5.	*Potato Hill andesite intercalation:* dark reddish-brown vesicular andesite porphyry in lower 25 feet; flow breccia, much weathered in upper 20 feet	45
4.	Tuff; gray to red, fine- to medium-grained	20
3.	*Sheep Canyon basalt intercalations:*	
	b. Fine-grained, dark basalt flow with vesicular top	32
	a. Fine-grained basalt flow with vesicular top	33
2.	Tuff, gray, fine- to medium-grained	11
1.	*Crossen trachyte intercalation:* porphyritic trachyte flow breccia, base of section	
	Total thickness measured	195

In Walnut Draw graben the brecciated phase is well exposed in the bed of the creek and in a fault slice that hangs on the trachyte forming the east wall of the graben. In this exposure 15–20 feet of andesite breccia is overlain by weathered andesite grading up to tuff which is overlain by the basal flow of the Cottonwood Spring basalt (Pl. 5, fig. 1).

Petrography.—Fine-grained hematite developed by oxidation of the primary ferromagnesian minerals obscures the minerals in the groundmass. The chief mineral is andesine

in small lathlike crystals, but there are smaller amounts of a more silicic feldspar, apatite, and magnetite-ilmenite. The large rounded and corroded phenocrysts form about 12 per cent

TABLE 6.—*Chemical analysis of Potato Hill andesite*
From Hill 4975 on Crossen Mesa
(S. S. Goldich, analyst)

SiO_2	61.46
Al_2O_3	14.85
Fe_2O_3	7.23
FeO	.68
MgO	.74
CaO	5.23
Na_2O	3.75
K_2O	2.58
H_2O+	.22
H_2O-	.17
CO_2	.10
TiO_2	1.52
P_2O_5	1.13
MnO	.11
BaO	.18
S	.00
	99.95
Sp. gr. (24°/4°)	2.764

of the rock. They are zoned and range from An_{55} in inner parts to An_{45} in the outer zones.

A large amount of secondary silica occurs as granular quartz filling vesicles and as fibrous chalcedony lining the walls of the cavities. Associated with the quartz and chalcedony is a fibrous mineral, probably celadonite, characterized by strong pleochroism—light yellow parallel to the length of the fibers and blue green at right angles. The brecciated andesite is similar in composition to the massive phase but is cemented with silica or calcite.

Chemical analysis.—The analyzed specimen (Table 6), from a point 1.4 miles northwest of Potato Hill, represents the lower massive phase of the porphyritic andesite. The high silica content reflects the large amount of introduced quartz and chalcedony, and the calculated quartz is 21 per cent of the norm (Table 10). The oxidation of the rock is revealed in the large percentage of ferric iron, and hematite is 7 per cent of the norm. Apatite is 2.7 per cent of the normative minerals. In other respects the analysis is not unusual.

Stratigraphic relations.—The Potato Hill andesite, like the Crossen trachyte and the Sheep Canyon basalt, is an intercalation in the Pruett tuff. The flow was weathered before being covered with volcanic ash, and as a result the contact between the andesite and the overlying tuff is not sharp. This tuff ranges from a few feet to 30 feet thick and is not differentiated on the geologic map (Pl. 1) which shows the Cottonwood Spring basalt resting on the Potato Hill andesite.

Cottonwood Spring Basalt

Definition and description.—The basalt flows lying on the Pruett formation and overlain by the Duff tuff are named for Cottonwood Spring which issues from the base of the flows in a deep gulch cut into the basalt-capped mesa south of Mitchell and Crossen mesas and north of the 02 ranch road. In this area the Cottonwood Spring basalt covers about 17 square miles. The flows continue westward in the Jordan Gap quadrangle where they cover a somewhat smaller area. Remnants of the flows cap small hills north and south of the 02 ranch house along the western boundary of the quadrangle. The basalt dips about 2° to the northwest passing under the Duff tuff, but in the northern part of the quadrangle, on the upthrown side of the Walnut Draw fault zone, the flows cap the hills that rise above the Crossen trachyte and occur in the fault blocks on the western edge of Crossen Mesa. Originally the Cottonwood Spring basalt formed a sheet covering a large part of the quadrangle and adjacent regions to the north and west.

The number of flows varies, and little significance is attached to this feature. Locally small fronts apparently pushed out ahead of the main body of lava to be overridden later. The lower parts of the flows commonly are flow breccias; the upper parts are vesicular and amygdaloidal with numerous fillings of quartz, chalcedony, jasper, and calcite. Large crystals of calcite showing unusual crystal forms and the rarely developed basal pinacoid were collected from cavities in the flows on the hill south of the 02 ranch house, but the crystal faces are etched and are not suitable for measurement. Pahoehoe or ropy lava is found in many places. The massive dense phase of the basalt is grayish black to dark

greenish black; the vesicular rock commonly is reddish gray to reddish brown. The flow breccia usually is composed of dense black to fine-grained vesicular basalt. Weathering apparently is controlled by topography. In gullies, low places, and on gentle slopes spheroidal forms develop; whereas on steep slopes and in cliffs a thin slabby structure results. The weathered basalt favors the growth of ocotillo.

The Cottonwood Spring basalt flows are thickest in the hills and mesas west of Whirlwind Spring where a maximum of 325 feet was measured. The original thickness is not known because in this area the Duff tuff has been stripped. The flows are divided into lower and upper units by a thin bed of tuff. The lower unit comprises two to six flows, totaling 100–150 feet. The upper unit in most places is a single massive flow or remnant of a thick flow which southwest of Cottonwood Spring is 164 feet thick, but in other places this thickness may represent five or more flows. The tuff bed breaking the succession of Cottonwood Spring basalt is 1–19 feet thick. Generally it is less than 5 feet thick and is baked, hard, dense, red rock. In thicker beds the tuff away from the contacts is gray and less indurated. Locally rounded fragments of vesicular basalt are contained in the lower part of the tuff bed, and in places the tuff was fragmented and incorporated in the base of the overriding flow.

In the type section (Appendix, Section No. 5) measured up the south point of the basalt mesa southwest of Cottonwood Spring the flows aggregate 300 feet. The lowermost flow resting on red tuff of the Pruett formation is 40–50 feet thick. The contact is marked by a resistant ledge, 3–4 feet thick, of coarse-grained basalt. The upper 9 feet is vesicular with many fillings of quartz, chalcedony, agate, and calcite. Four flows ranging from 6 to 15 feet thick and one 50 feet thick complete the lower unit of 130 feet of basalt in six flows beneath a tuff bed, 8–19 feet thick. Above the tuff is 164 feet of basalt, apparently a single thick flow with a prominent flow breccia of vesicular basalt at the base and a 20-foot amygdaloidal top. This flow caps the mesa and makes cliffs 60 to 80 feet high on the southern scarp. The thickness of amygdaloid suggests that erosion has not removed much of the flow, but younger flows probably were eroded, and the original thickness may have been greater than the 300 feet of basalt measured.

A good section of the Cottonwood Spring basalt is exposed on Hill 4631 (Fig. 6) 1½ miles northeast of Cottonwood Spring where nine flows are 300 feet thick. The lower part of the section beneath a 7-foot layer of red tuff is 154 feet thick and consists of four flows of which the lower one makes up 83 feet. Above the tuff are five flows which aggregate 148 feet. Southwest of Hill 4321 in the spur at Whirlwind Spring a similar succession of basalt flows with separating tuff bed is 325 feet thick.

At Potato Hill on Crossen Mesa the basal Cottonwood Spring flow is 85 feet thick; a total thickness of 166 feet includes six flows to the top of the hill. North of Potato Hill in Hill 4806, only 67 feet of the basal flow remains. In Hill 4975 (Fig. 6) there are six flows. Five flows beneath a 7-foot tuff bed total 110 feet of which the basal flow makes up 61 feet. The remnant of a thick flow above the tuff is 55 feet thick making the succession, including the tuff, 172 feet in total thickness. The upper flow of Hill 4975 also caps Hill 5016 three-fourths of a mile to the northwest where 135 feet of this flow remains. Immediately below is a dense red tuff, 8 feet thick, which is underlain by 92 feet of basalt in two flows. The total thickness of 235 feet is the maximum measured for the Cottonwood Spring basalt in the northern part of the quadrangle.

In Walnut Draw graben the Cottonwood Spring flows are overlain by the Duff tuff. On the south slope of the elongated hill in the graben 175 feet of basalt was measured, and on the basis of amygdaloidal tops there are seven flows below a thin layer of red tuff. Above the tuff is 80 feet of basalt which caps the hill and forms prominent cliffs. The base of the section is covered by alluvium, but a short distance to the south the Potato Hill andesite is exposed in the bed of Walnut Draw.

Petrography.—Thin sections show a range in composition from trachyandesite or latite to olivine basalt, but no detailed studies of variations between or within the flows have been made. The flows typically are fine- to

medium-grained, ophitic to diabasic, uniform and nonporphyritic or with few small phenocrysts. Flow structure, fractured and broken crystals, and wavy extinction are characteristic. Some of the reddish-brown flows are highly oxidized, and the original ferromagnesian minerals including much of the magnetite have been altered to or obscured by hematite. Other secondary minerals are analcime, serpentine, uralitic amphibole, chlorite, and calcite.

Thin sections of the analyzed specimen (Table 7, No. 2) from the basal flow at the type locality southwest of Cottonwood Spring contain 60–70 per cent of labradorite. The plagioclase, rarely showing well-developed twinning bands, is zoned and commonly has a mantle of orthoclase. Orthoclase also occurs in clear fresh grains together with granular aggregates of analcime interstitial to the plagioclase laths. Olivine, 20–25 per cent of the rock, is largely altered to serpentinous products that penetrate the near-by fractured grains of plagioclase. Pale-green fibrous chrysotile with weak birefringence, and dark-green to yellow and brown iddingsite or bowlingite with stronger birefringence, good cleavage, and marked pleochroism are common. Pyroxene rarely exceeds 10 per cent of the rock and is normal augite, some with a purplish tinge. Most of the augite is fresh, but some is altered to long slender needles of actinolite and to chlorite. Magnetite and apatite are abundant.

Analcime occurs in minor amounts in clear grains between the labradorite laths and in dusty granular aggregates replacing the plagio-

TABLE 7.—*Chemical analyses of the basal flow of the Cottonwood Spring basalt, Butcherknife basalt, and similar rocks*

	1	2	3	4	5	6
SiO_2	46.90	48.08	48.53	49.80	46.87	44.81
Al_2O_3	15.82	15.87	17.42	18.38	17.04	16.81
Fe_2O_3	4.18	3.74	3.46	1.84	3.12	4.06
FeO	7.74	7.49	7.48	7.57	6.81	7.19
MgO	4.48	3.86	3.10	2.63	3.35	4.21
CaO	6.24	5.76	6.60	7.09	7.66	7.21
Na_2O	3.80	4.50	5.13	4.94	5.13	4.65
K_2O	2.32	2.43	2.38	2.60	2.39	1.97
H_2O+	1.93	2.17	1.68	1.73	2.99	3.70
H_2O-	1.13	.97	.26	.27	.21	1.07
CO_2	.02	.02	.02	—	—	—
TiO_2	2.79	2.68	2.53	2.30	2.96	3.27
P_2O_5	1.97	1.81	1.10	.95	.93	1.04
MnO	.21	.23	.21	—	.19	—
BaO	.23	.12	.06	—	—	—
SrO	.10	.09	.13	—	—	—
S	.00	.00	.01	—	—	—
	99.86	99.82	100.10	100.10	99.65	99.99
Sp. gr.	2.823	2.762	2.808			

1. Cottonwood Spring basalt, basal flow at Potato Hill on Crossen Mesa. Cat. No. 817. Sp. gr., 26°/4°.
2. Cottonwood Spring basalt, basal flow, three-fourths of a mile southwest of Cottonwood Spring. Cat. No. 840. Sp. gr., 24.5°/4°.
3. Basalt at Butcherknife Hill. Cat. No. 830. Sp. gr., 23.4°/4°.
4. Analcite trachybasalt porphyry, Mountain 3513, Terlingua-Solitario region, Texas.
5. Analcite syenogabbro, Cigar Mountain, Terlingua-Solitario region.
6. Analcite trachydolerite, 1 mile northeast of Mountain 3513, Terlingua-Solitario region.

Nos. 1, 2, and 3, S. S. Goldich, analyst.
Nos. 4, 5, and 6, R. B. Ellestad, analyst (Lonsdale, 1940, p. 1620).

clase. Its abundance is difficult to estimate, about 5 per cent of the rock. Likewise orthoclase is in sufficient amount to justify modification of the field name basalt, and the analyzed specimen is classified as analcime trachybasalt following the usage of Lonsdale (1940) for similar rocks in the Terlingua-Solitario region to the south.

Chemical analysis.—Analyses (Table 7) of samples of the basal flow of the Cottonwood Spring basalt from Potato Hill and from near Cottonwood Spring are similar in most respects. The samples also resemble analysis No. 3 of the dense basalt from Butcherknife Hill and the analyzed analcime-bearing rocks from the Terlingua-Solitario region (Lonsdale, 1940, p. 1620). There are minor differences between the analyzed samples of the Cottonwood Spring basalt and the Sheep Canyon basalt flows (Table 5), but it is not likely that the two groups of flows can be separated by chemical analyses of scattered samples. Selected samples probably would show a greater chemical diversity within each group than between them.

Stratigraphic relations.—On Crossen Mesa the Cottonwood Spring basalt rests on tuff deposited on the weathered surface of the Potato Hill andesite. The tuff at Potato Hill is 25 feet thick, and the fossil soil on the andesite is thin; elsewhere the tuff is thin and the weathered zone is as much as 20 feet thick. The weathered andesite grades upward to red tuff baked to a dense hard rock at the time of extrusion of the basal Cottonwood Spring flow. South of Crossen Mesa beyond the southern limit of the andesite, the Cottonwood Spring basalt rests on the Pruett tuff which immediately below the basal flow is a hard red rock, but below this baked layer is 30 feet of soft, earthy, red material which probably represents tuff weathered contemporaneously with the andesite.

The Cottonwood Spring flows were partly eroded before deposition of the Duff tuff, and in part this erosion accounts for the thinning of the flows from a thickness of over 300 feet near Whirlwind Spring to less than 50 feet 10 miles west in the Jordan Gap quadrangle. The flows are variable in number and in thickness, but both on Crossen Mesa and in the basalt mesa to the southwest the basal flow is relatively thick, 43–85 feet. In both areas the succession of flows with included tuff bed is similar, and there is little doubt that the flows are the same and that they were originally poured out on a relatively flat and probably weathered surface of the Pruett tuff (Figs. 3, 6).

Duff Formation

Definition and description.—The Duff formation, named for Duff Springs in the northwestern part of the Buck Hill quadrangle, is chiefly rhyolitic tuff with minor breccia and conglomerate. The tuff lies on the Cottonwood Spring basalt and is overlain by the Mitchell Mesa rhyolite. In addition to the outcrops in the Mitchell Mesa area there are two small outcrops on the small elongated hill south of the 02 ranch house. The Duff formation covers most of the Jordan Gap quadrangle to the west.

The Duff tuff is predominantly light gray and in strong sunlight is dazzling white in contrast to the darker color of the flows and the duller somber hues of the Pruett tuff. Light shades of red and yellow are common, and alternations of colors give some of the Duff a variegated appearance. Most of the material is fine-grained, well-indurated ash grading locally to tuff-breccia and breccia-conglomerate. Much of the tuff is massive; some is cross-bedded; and a large part is well stratified, ranging from thin laminations to beds 2–6 inches thick, but beds 1 foot or more thick are common. The tuff weathers to chalk-white nodular forms, and the gentler slopes are strewn with pellets and small rounded pebbles of the tuff.

Lenticular beds of conglomerate which generally do not persist for any great distance occur throughout the formation, but probably are more numerous in the lower part. Conglomerate beds are fairly persistent at two levels, however. The lower level is approximately 40 feet above the base of the formation; the upper level is 200–250 feet below the top. These conglomerate beds attain 40 feet in thickness and are characterized by cross-bedding and channeling. They are generally dark brown and stand out in the light-gray tuff.

The Duff formation is 1015 feet thick in the type section measured from the outcrop of the Cottonwood Spring basalt 1¼ miles northwest of Duff Springs up a large gully on the southern slope of Mitchell Mesa to the rhyolite caprock. Along the line of section the lower conglomerate, 6 feet thick, is 37 feet above the base of the formation. Two beds of conglomerate near the top of the formation are separated by 46 feet of tuff. The lower bed with igneous boulders 1 foot or more in diameter is 8 feet thick; the upper bed consisting mainly of pebbles and cobbles is 10 feet thick and is 200 feet below the base of the Mitchell Mesa rhyolite. Graham (1942) measured the Duff formation at Whirlwind Mesa in the Jordan Gap quadrangle, approximately 4 miles west of Mitchell Mesa. At this locality the Cottonwood Spring basalt is covered by alluvium, but, including the lower conglomerate which is exposed, the Duff section is 1020 feet thick. The lower conglomerate is 31 feet thick, and an upper single bed of conglomerate, 250 feet below the top of the formation, is 38 feet thick. Allowing 40 feet for the covered interval below the conglomerate at the base of the section, the total thickness of the Duff formation is 1060 feet. At McKinney Mountain 10 miles south of the Whirlwind Mesa section, Graham (1942) measured a thickness of 1400 feet for the Duff formation. The tuff was deposited on an irregular, eroded surface of the Cottonwood Spring basalt, and a variable thickness is expected. The maximum thickness may exceed the measured thickness at McKinney Mountain.

Numerous dikelike masses cut the Duff tuff. These are clastic dikes composed of indurated clay or brecciated tuff, and slickensides and brecciated textures indicate that they are the result of small movements probably related to the faulting. Water moving along the brecciated zones deposited crystals of analcime and calcite in fractures. The clastic dikes are generally vertical or at high angles, and because they are more resistant than the normal tuff erosion leaves wall-like remnants or natural stone fences (Pl. 3, fig. 1).

Composition.—Unlike the Pruett tuff, the Duff rarely is calcareous. A small amount of calcite commonly is found near the clastic dikes and appears to be related to water movement along the dikes. The analyzed sample (Table 2, No. 2) of the Duff tuff is a well-indurated, light-gray, fine-grained, rhyolitic ash in which numerous small flakes of biotite are recognizable with a hand lens. The sample was collected from the east scarp of Mitchell Mesa 100 feet below the top of the formation. In most respects the analysis of the Duff sample is similar to that of the Pruett tuff, but its relative content of alkalies is somewhat greater, and its percentage of lime is considerably less. The Duff tuff, like the Mitchell Mesa rhyolite (Table 8, No. 1), is potassic.

The rhyolitic composition of the tuff also is revealed in petrographic studies of 18 samples collected by Graham (1942) from the formation at McKinney Mountain. The index of refraction of the volcanic glass in samples fairly regularly spaced from bottom to the top of the formation is 1.492 ± .003, indicating a silica content of approximately 73 per cent. Heavy-liquid separations made on pulverized samples, from which the very fine material was removed by washing, yielded uniformly small fractions of heavy minerals, rarely as much as 0.5 per cent by weight of the washed sample. The light fractions, in addition to volcanic glass and its alteration products, contained abundant feldspar and quartz. Magnetite-ilmenite and alteration products are the chief heavy minerals. Biotite is the most abundant nonopaque mineral and occurs in all samples. There are two varieties of biotite. One displays a yellowish-green to dark olive-green pleochroism; the other is a deep reddish-brown variety. Apatite and zircon follow in order of abundance. The apatite is chiefly in subrounded grains, the zircon in well-formed crystals. No variation in relative abundance of biotite, apatite, or zircon with stratigraphic position was found. Less important than these minerals both in relative abundance and in frequency of occurrence are aegirine-augite, hornblende, basaltic hornblende, epidote, sphene, rutile, and muscovite. Aegirine-augite is rare in the samples from the lower half of the formation but appears to be common in the upper part. It is abundant in well-rounded grains in a sample taken near the top of the section.

A variety of volcanic and sedimentary rocks is represented in the conglomerate beds which contain abundant dense and vesicular basaltic types and dense to porphyritic light-colored aphanites such as rhyolite and trachyte. Locally pebbles of fine-grained syenite are common. The sedimentary pebbles are largely chert and fine-grained limestone. Coarse sand commonly is associated with the upper conglomerate. Sand and tuffaceous material cement the conglomerates.

Stratigraphic relations.—In the Buck Hill quadrangle and in the northern part of the Jordan Gap quadrangle the Duff tuff is easily delimited by the Cottonwood Spring basalt at the base and by the Mitchell Mesa rhyolite at the top. In view of the thinning of the Cottonwood Spring flows, the Duff tuff in places may have been deposited directly on the Pruett tuff. Such a condition has not been recognized to date in the areas studied.

Mitchell Mesa Rhyolite

Definition and description.—The thin sheet of rhyolite capping Mitchell Mesa in the northwestern corner of the Buck Hill quadrangle is named for the mesa. The rhyolite resting on the Duff tuff covers about 3 square miles of Mitchell Mesa and is an important lithologic unit in this region as the caprock of the high mesas.

The rhyolite is easily recognized by its pink color and prominent glassy rounded phenocrysts of quartz and opalescent tabular phenocrysts of sanidine. The phenocrysts, rarely exceeding 0.2 inch in length, are set in a dense pink or reddish-gray groundmass. The uniform texture of the rock is broken by light-gray vesiculated areas and by red inclusions of baked and silicified ash. The fracture is subconchoidal to irregular. The weathered rock is dark reddish gray (rusty) or black. Unlike the pitted surface of weathered Crossen trachyte, the weathered surface of the rhyolite commonly shows numerous small protuberances of the more resistant phenocrysts.

The rhyolite on the southern edge of Mitchell Mesa is 38 feet thick and probably does not exceed 50 feet on the mesa. A similar thickness caps Whirlwind Mesa just west in the Jordan Gap quadrangle, but Graham (1942) measured 70 feet of rhyolite on the rim of Antelope Mesa 10 miles southwest of Mitchell Mesa and 150 feet on McKinney Mountain.

Petrography.—The fine-grained groundmass consists of quartz, alkalic feldspar, and unidentified material which may be rhyolitic glass and its devitrification products. Magnetite, zircon, and apatite are accessory minerals. Alteration products include small amounts of limonite, hematite, and leucoxene. The rock owes its pink color to finely disseminated hematite. There is some silica (chalcedony) of late introduction. The phenocrysts are quartz and sanidine, rounded and embayed by magmatic resorption. Optical data for the feldspar are $\alpha = 1.522$, $\beta = 1.528$, $\gamma = 1.528$; all $\pm .002$. $(-)$ 2V is small. The optic plane is normal to 010, and Z = b; $X \wedge a = 5° +$. The dispersion is $r > v$.

Chemical analysis.—A chemical analysis of the Mitchell Mesa rhyolite and analyses of other rhyolitic rocks from Trans-Pecos Texas are given in Table 8. The high percentage of potash (6.34 per cent K_2O) in the analyzed sample of Mitchell Mesa rhyolite is noteworthy, because the analyses of rhyolitic rocks from Trans-Pecos Texas indicate these rocks are sodi-potassic. The closest approach to the composition of the Mitchell Mesa rhyolite is an analysis of a potassic rhyolite from the Solitario (Table 8, No. 2). This analysis, however, appears to be of a weathered specimen, and because soda undergoes greater losses than potash by weathering the analysis may deviate considerably from the composition of the fresh rock. The analyses of rocks from near Shafter (Fig. 2) and from west of Paisano Mountain (Table 8, Nos. 3, 4) are typical of available analyses of rhyolitic rocks in this part of Texas.

Stratigraphic relations.—The Mitchell Mesa rhyolite is the youngest rock of the volcanic series in the Buck Hill quadrangle. In Goat Mountain, 6 miles north in the Alpine quadrangle, and in the Jordan Gap quadrangle younger unnamed tuff beds and flows occur above the rhyolite. Reconnaissance work in these areas indicates the Mitchell Mesa rhyolite is a flow.

TABLE 8.—*Chemical analyses of Mitchell Mesa rhyolite and other rhyolites from Trans-Pecos Texas*

	1	2	3	4
SiO_2	78.83	76.23	75.12	72.86
Al_2O_3	9.95	12.11	10.94	11.74
Fe_2O_3	1.40	.42	2.88	2.71
FeO	.10		.86	1.66
MgO	.16	.46	.07	.06
CaO	.35	.36	.20	.24
Na_2O	1.78	1.40	4.46	4.63
K_2O	6.34	4.93	4.54	4.92
H_2O+	.41	1.47	.19	.40
H_2O-	.11	1.80	.18	.51
CO_2	.27	—	.04	tr.
TiO_2	.20	.13	.20	.20
P_2O_5	.03	.01	none	tr.
MnO	.08	—	.08	.07
BaO	tr.	—	none	none
SrO	n.d.	—	none	none
Li_2O	n.d.	—	tr.	tr.
Rare earths	n.d.	—	.04	.05
SO_3	n.d.	—	none	—
S	tr.	—	.05	tr.
ZrO_2	n.d.	—	.13	.28
Ignition loss	—	.92	—	—
	100.01	100.24	99.98	100.33
Sp. gr.	2.596	—	2.617	2.635

1. Rhyolite porphyry capping Mitchell Mesa. Cat. No. 599. Sp. gr., 23.4°/4°. S. S. Goldich, analyst.
2. Potassic rhyolite, East Tank, Solitario. Phenocrysts of quartz and orthoclase. Groundmass of alkali feldspar and quartz. R. B. Ellestad, analyst (Lonsdale, 1940, p. 1620).
3. Rhyolite near Shafter, Shafter quadrangle. "*Liparose near alaskose.* Phenocrysts of sanidine and quartz. Groundmass of alkali-feldspar, quartz, riebeckite (?), and aegirite. Spherulitic bands traverse the rock. Sp. gr., 2.617, 15.5°." W. F. Hillebrand, analyst (Clarke, 1904, p. 74).
4. Rhyolite west of Paisano Mountain, Alpine quadrangle. "*Liparose.* Contains alkali-feldspars, quartz, arfvedsonite, and aegirite. Sp. gr., 2.635, 15.5°." W. F. Hillebrand, analyst (Clarke, 1904, p. 74).

Volcanic Section in Elephant Mountain

Principal features.—On the southern slope of Elephant Mountain a considerable thickness of the Pruett tuff is overlain by flows of the Sheep Canyon basalt. The flows pinch out to the north, and the massive syenite which caps Elephant Mountain in the Alpine quadrangle lies on both the basalt and the tuff (Pl. 6). In the Buck Hill quadrangle the southern end of the mountain is downthrown along northwest-trending faults. On the northeast side of the larger fault the Cretaceous formations are exposed, and the Boquillas is overlain by the Pruett. Southwest of the fault the Cretaceous does not crop out, and the lower Pruett beds are covered by alluvium.

Pruett formation.—The Pruett formation in the Elephant Mountain area is 500–600 feet thick, of which approximately 300 feet is described in the section (Appendix, Section No. 9) measured in the large gully on the southern slope of the mountain. The basal beds are tuffaceous conglomerate and sandstone or light-yellow to gray sandy tuff which weathers to a characteristic bright yellow. These beds are succeeded by varicolored, but dominantly gray, tuff. In the lower part of the section the tuff is massive with indistinct stratification. It is well lithified and weathers to pebbly nodular gravel. There are a few minor lenses of breccia and conglomerate. In the upper part sorting and stratification are better developed and thin layers of calcareous tuff and freshwater limestone interfinger with the tuff.

Sheep Canyon basalt.—There are at least three thick flows of the Sheep Canyon basalt in the southern slope of Elephant Mountain. Along the line of section (Appendix, Section No. 9) the lowest flow is 88 feet thick with an amygdaloidal top 25–30 feet thick. The lower part of the flow is dense porphyritic basalt. Light-gray to brown cherty limestone, 4–13 feet thick and containing silicified fresh-water gastropods, lies between the first flow and a middle porphyritic basalt flow 121 feet thick. Plagioclase phenocrysts, as much as 1 inch long, are contained in a dark fine-grained groundmass. The upper 20–30 feet of the flow is vesicular with numerous chalcedony, jasper, and calcite fillings. A bed of calcareous tuff with lenses of limestone generally separates the middle from the upper flow, but is missing along the line of section. This calcareous bed usually is 5 feet and locally as much as 15 feet thick. The upper basalt flow is 171 feet thick, but an interval of 50 feet above the highest exposure is covered, and the amyg

daloidal zone expected at the top of the flow was not found, so the thickness of 380 feet measured for the flows probably does not represent the total thickness. A probable maximum thickness is 450 feet. The Sheep Canyon basalt flows thin rapidly northward, and on the east side of Elephant Mountain, 1¼ miles north of the Buck Hill quadrangle, the flows are absent (Pl. 6). A similar pinch-out occurs on the west side of the mountain 2½ miles north of the Buck Hill quadrangle.

An analysis of the porphyritic basalt from Elephant Mountain (Table 5, No. 3) closely resembles the porphyritic Sheep Canyon basalt on Crossen Mesa. The Elephant Mountain sample was collected in a large gully near the base of the basalt, 105 feet thick, overlain by 10 feet of light-gray fresh-water limestone beneath the syenite. Heavy talus obscures much of the outcrop, and the field observations are inconclusive as to which of the three Elephant Mountain flows is represented. The sample was selected for analysis because it represents the freshest material collected. Approximately 1000 feet north of this locality the basalt feathers out, and the syenite rests on limestone and tuff of the Pruett formation.

Stratigraphic relations.—The tuff, conglomerate, sandstone, and fresh-water limestone beds in Elephant Mountain are part of the Pruett formation. Large low-spired gastropods collected from the lower part of the section are similar to those collected from the ash at the base of Crossen Mesa west of Calamity Creek. Fresh-water limestone beds in the tuff and between the basalt flows in the Elephant Mountain area closely resemble the limestone in the Sheep Canyon area west of Calamity Creek (Pl. 6). In both areas *Goniobasis* is common in the limestone. North of Elephant Mountain the tuff beds are overlain by the Crossen trachyte. C. L. Baker (personal communication, 1941) found tuff, calcareous tuff, and fresh-water limestone, 500 feet thick capped with a trachyte flow 120 feet thick, resting on the Cretaceous limestone on the western flank of Mt. Ord 12 miles north of Elephant Mountain.

In summary the following points support the correlation of the volcanic rocks in Elephant Mountain with the Pruett-Sheep Canyon basalt sequence west of Calamity Creek: (1) The stratigraphic succession of tuff, calcareous tuff, fresh-water limestone, and basalt flows in two areas is similar. (2) Fossils in the fresh-water limestone in the tuff and between the basalt flows are similar. (3) The basalt flows in the two areas are similar chemically and are porphyritic; they are the only porphyritic basalts known in the region. (4) The flows in both areas occupy former stream channels which are part of the ancient drainage system developed after extrusion of the Crossen trachyte and before extravasation of the Sheep Canyon basalt.

Quaternary Alluvium

Alluvial deposits cover large areas in the Buck Hill quadrangle and range from clay and silt to boulders. The deposits are chiefly basin fill which accumulated in Green Valley to a thickness of 30 feet or more. Calamity Creek, The Ditch, and Terlingua Creek in many places are steep-walled arroyos which have cut through the alluvium into the bedrock.

Albritton and Bryan (1939) on the basis of disconformities have divided the alluvial deposits in the valley flats of the southern Davis Mountains into three formations, from oldest to youngest: (1) Neville, (2) Calamity, and (3) Kokernot. In the present investigation the alluvium is not differentiated.

Intrusive Igneous Rocks

General Statement

Syenitic igneous rocks with definite intrusive relationships form two prominent hills in the southern part of the quadrangle. In the south-central part syenite forms the flat-topped Buck Hill for which the quadrangle is named. West of Buck Hill a syenite plug is the core of Straddlebug Mountain.

Approximately 4 miles north of Buck Hill a small mass of dense black basalt forms Butcherknife Hill.

Microsyenite

Syenite of Buck Hill.—The syenite of Buck Hill is a light-gray, fine-grained, even-textured

rock weathering to yellowish brown. The tabular mass is cut by two normal faults that divide it into three segments with the middle block downthrown. The largest and southwestern segment measures 4500 feet in length in a northwest-southeast direction and 3000 feet across at its widest part. The middle segment is 2500 feet wide at its southeastern end, but splits to the northwest into a number of narrow offshoots. The northeastern segment is about 2000 feet long and 700 feet wide. The greatest overall dimension of the syenite mass is 6500 feet along an east-west line.

The syenite is a small intrusive which apparently spread sill-like at or near the contact of the thin-bedded Boquillas limestone and the overlying basal beds of the Pruett. The mass averages 65 feet thick, but it may have been thicker as its roof has been removed. In the northeastern segment thin-bedded Boquillas limestone is sandwiched between massive sheets of the syenite (Pl. 5, fig. 2). On the southeastern side of the hill the middle segment of the syenite overlies a thin bed of the basal Pruett conglomerate which rests on the Boquillas limestone. The conglomerate was not found on the northwestern side of the middle segment where narrow dikes or offshoots from the main syenite mass cut the Boquillas. One of these apophyses is over 2000 feet long. In the southwestern segment, which is the upthrown side of the larger fault, the syenite lies on a considerable thickness, 14–30 feet, of the Pruett formation. Southeast of Bench Mark 4203 there are xenoliths of the Pruett sandstone in the syenite.

Syenite of Straddlebug.—The syenite of Straddlebug Mountain closely resembles that of Buck Hill, but it is more weathered. The small plug measures less than 1000 feet across, but, unlike the Buck Hill intrusive, it resulted in a pronounced deformation and doming of the sedimentary rocks. The oldest beds brought to surface are Buda limestone; the youngest are lower beds of the Boquillas. The sedimentary rocks dip away from the hill on all sides except on the north where they are covered by alluvium on the downthrown side of the Chalk Draw fault. Dips as high as 80° were observed on the southwest side of Straddlebug, but usually they are considerably less. At the contact with the syenite the Boquillas flags have been baked to dark, fine-grained marble. On the southwest side of Straddlebug Mountain a sill of syenite not more than 6 feet thick intrudes the Boquillas and appears to have had little effect on the limestone. The sill is estimated to be about 50 feet above the base of the Boquillas formation.

Petrography.—The freshest specimens of syenite from Buck Hill show some weathering, and in samples from Straddlebug Mountain the weathering is pronounced. The feldspars are dusty, and the ferromagnesian minerals are altered to limonite. The texture is fine-grained, trachytic to granular. Extinction in the feldspar laths is anomalous, and the twinning is indistinct. Large anhedral grains of feldspar appear to be a microperthitic intergrowth. Nepheline was not identified, and zeolitic material is rare. In the syenite of Buck Hill there is some hornblende with deep olive-green to brown pleochroism and a small amount of biotite with light-yellow to reddish-brown pleochroism. Magnetite and apatite are accessory. The intrusives are alkalic microsyenite.

Chemical analysis.—A chemical analysis of a sample of syenite from Buck Hill is given in Table 9 in which three analyses of similar syenites are included. Ferric and ferrous oxides and some of the minor constituents were determined on a sample of the Straddlebug syenite. The relatively large proportion of ferric iron is indicative of the weathering of these masses. The percentage of alkalies is large, and calculated normative nepheline is 2 per cent of the Buck Hill rock (Table 10).

Age and correlation.—Because of similarity in composition and because both the Straddlebug Mountain and Buck Hill intrusives are on an east-west line of structural weakness, the two intrusives may be related and of the same age. The syenite of Buck Hill is clearly younger than the basal conglomerate of the Pruett formation.

Many syenitic intrusives in this region closely resemble the microsyenite of Buck Hill and Straddlebug Mountain. Two conspicuous masses are the syenite on Elephant Mountain and the Santiago Peak intrusive east of the quadrangle. The Elephant Mountain syenite

TABLE 9.—*Chemical analyses of syenite from Buck Hill and other localities in Trans-Pecos Texas*

	1	2	3	4	5
SiO_2	60.98	—	62.46	59.36	59.62
Al_2O_3	17.35	—	17.10	18.20	17.89
Fe_2O_3	5.46	5.25	2.49	6.19	3.97
FeO	.49	.08	2.65		1.83
MgO	.33	—	.28	.15	.75
CaO	1.38	—	1.27	1.64	2.78
Na_2O	5.98	—	6.84	5.99	6.20
K_2O	5.58	—	5.44	5.28	4.55
H_2O+	.81	—	.49	.22	.71
H_2O-	.26	—	.15	1.10	.38
TiO_2	.50	—	.38	tr.	.66
ZrO_2	.16	—	.10	—	—
CO_2	.32	—	tr.?	—	—
P_2O_5	.28	.10	.11	tr.	.26
MnO	.21	.18	.18	tr.	—
BaO	.03	tr.	none	—	—
SrO	n.d.	—	none	—	—
Li_2O	n.d.	—	tr.	—	—
S	.00	.00	none	—	—
Cl	n.d.	—	n.d.	.31	—
Rare earths	n.d.	—	.03	—	—
	100.12		99.97	98.44	99.60

1. Alkalic microsyenite of Buck Hill. Sp. gr., 25°/4°, 2.659. S. S. Goldich, analyst.
2. Alkalic microsyenite of Straddlebug Mountain. S. S. Goldich, analyst.
3. Pulaskite, Santiago Mountain, Santiago Peak quadrangle. Sp. gr., 2.581, 25.5°. W. F. Hillebrand, analyst (Clarke, 1904, p. 75).
4. Alkalic syenite, Elephant Mountain, Alpine quadrangle. L. A. Mikeska and W. A. Felsing, analysts (Schoch, 1918, p. 185).
5. Analcitic microsyenite, Sawmill Mountain, Terlingua-Solitario region. R. B. Ellestad, analyst (Lonsdale, 1940, p. 1620).

rests on tuff and fresh-water limestone of the Pruett formation and on flows of the Sheep Canyon basalt. The Elephant Mountain syenite (Hardin, 1942) ranges from 800 to 1100 feet in thickness, but some of the upper part may have been removed by erosion. In the valley just north of Elephant Mountain on the Neville ranch three small intrusives of similar syenite pierce both the Boquillas limestone and the basal Pruett tuff (Pl. 6). These probably are the feeders for the Elephant Mountain syenite which is considered to be intrusive. The time of intrusion is post-Sheep Canyon basalt and probably post-Duff.

In the Santiago Peak quadrangle Eifler (1943, p. 1635) found Cenozoic sandstones and conglomerates on the eroded Boquillas surface which on the southern and western sides of Black Mountain are overlain by syenite capping the mountain. At Santiago Peak the sandstone is overlain by white and lavender tuffs 800–900 feet thick. The syenite of Santiago Peak towers 1200 to 1300 feet above the highest level at which the tuff was found. These figures are significant because they indicate the Santiago Peak syenite probably was intruded into a sequence with a minimum thickness of the Buck Hill volcanic series. The sandstone and conglomerate may be the equivalent of the basal conglomerate member of the Pruett formation, and the total thickness of sandstones and tuff corresponds closely to the thickness of the Pruett. Originally deposits equivalent to the combined thickness of the Pruett and the Duff probably covered the region, and into these deposits the syenitic magma of Santiago Peak was intruded.

The syenite rocks of Santiago Peak and of Elephant Mountain are mineralogically similar and on a line connecting these masses are other large syenitic intrusives in the Santiago Peak quadrangle. These intrusives, the syenite of Buck Hill and of Straddlebug Mountain, and a large number of others in this region probably are closely related genetically.

Butcherknife Basalt

General features.—The Butcherknife basalt is a fresh, black, very fine-grained rock with a conchoidal fracture. Because the rock spalls readily, the upper slopes of the hill are covered with basalt fragments, so the relations of the basalt to the sediments could not be determined. Near the base of the hill tuffaceous coarse sandstone, sandy and calcareous tuff, and gray ash with thin beds and nodules of limestone of the basal Pruett rest on Boquillas limestone. The Boquillas strata away from Butcherknife Hill are essentially flat-lying, but the Pruett beds closer to the basalt show local disturbance, and on the northwest side of the hill dip up to 25° NW. The high dips in the Pruett beds in the vicinity of Butcherknife

TABLE 10.—*Chemical analyses and norms of Buck Hill rocks*

(S. S. Goldich, analyst)

	Sheep Canyon basalt			Cottonwood Spring basalt		Butcher-knife basalt	Syenite of Buck Hill	Potato Hill andesite	Crossen trachyte	Mitchell Mesa rhyolite	Volcanic tuff	
	Hill 4321	Potato Hill	Elephant Mountain	Potato Hill	Cottonwood Spring						Pruett	Duff
ANALYSES												
SiO_2	44.33	45.65	45.97	46.90	48.08	48.53	60.98	61.46	71.17	78.83	69.73	70.57
Al_2O_3	14.69	16.48	17.26	15.82	15.87	17.42	17.35	14.85	12.64	9.95	12.06	12.19
Fe_2O_3	4.27	6.92	2.84	4.18	3.74	3.46	5.46	7.23	4.28	1.40	1.06	1.29
FeO	8.68	6.40	9.18	7.74	7.49	7.48	0.49	0.68	0.11	0.10	0.13	0.09
MgO	4.88	4.05	5.63	4.48	3.86	3.10	0.33	0.74	0.13	0.16	0.88	0.60
CaO	7.00	7.48	7.87	6.24	5.76	6.60	1.38	5.23	1.05	0.35	2.16	1.43
Na_2O	3.46	3.38	4.33	3.80	4.50	5.13	5.98	3.75	3.91	1.78	0.78	1.80
K_2O	2.15	1.42	1.00	2.32	2.43	2.38	5.58	2.58	5.05	6.34	4.17	5.09
H_2O+	2.15	2.10	1.57	1.93	2.17	1.68	0.81	0.22	0.23	0.41	5.05	3.36
H_2O-	1.09	1.39	0.20	1.13	0.97	0.26	0.26	0.17	0.12	0.11	3.31	2.84
CO_2	0.07	0.37	0.00	0.02	0.02	0.02	0.32	0.10	0.19	0.27	0.01	0.14
TiO_2	3.64	3.21	2.87	2.79	2.68	2.53	0.50	1.52	0.56	0.20	0.18	0.22
P_2O_5	2.99	0.87	0.96	1.97	1.81	1.10	0.28	1.13	0.07	0.03	0.04	0.08
MnO	0.23	0.18	0.17	0.21	0.23	0.21	0.21	0.11	0.14	0.08	0.01	0.01
BaO	0.17	0.09	0.07	0.23	0.12	0.06	0.03	0.18	0.08	tr.	0.13	0.04
SrO	0.14	0.09	0.10	0.10	0.09	0.13	n.d.	n.d.	n.d.	n.d.	0.09	n.d.
S	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	tr.	tr.	0.00	0.00
ZrO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.16	n.d.	n.d.	n.d.	n.d.	n.d.
Total	99.94	100.08	100.02	99.86	99.82	100.10	100.12	99.95	99.73	100.01	99.79	99.75
Sp. gr. (t°/4°)	2.818	2.829	2.874	2.823	2.762	2.808	2.659	2.764	2.648	2.596	2.368	2.421
NORMS												
Q							1.5		29.5	36.5	29.5	28.5
or	12.5	8.5	6.0	13.5	13.0	14.0	33.5	15.5	30.5	38.0	25.0	30.5
ab	29.5	28.5	36.5	32.0	38.0	43.5	50.5	31.5	33.0	15.0	6.5	15.0
an	20.0	23.0	22.5	20.5	19.5	19.5	6.5	19.5	5.5	1.5	10.0	6.0
ne												
c												
wo	5.5	6.5	3.5	3.5	2.5	3.5		3.5				
en	12.0	10.0	14.0	11.0	9.5	7.5	0.5	1.5				
fs	10.5	8.0	11.5	9.5	9.5	9.0	0.5	1.5				
mt	6.0	9.5	4.0	6.0	5.5	5.0	7.5	10.5	6.0	2.0	1.5	1.5
il	7.0	6.0	5.5	5.5	5.0	4.5	1.0	3.0	1.0	0.5	0.5	0.5
ap	7.0	2.0	2.5	4.5	4.0	2.5	0.5	2.5	0.5	0.5	0.5	0.5
cc	0.5	1.0	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
hm												
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

Total	99.94	100.08	100.02	99.86	99.82	100.10	100.12	99.93	99.73	100.01	99.79	99.73
Sp. gr. (t°/4°)	2.818	2.829	2.874	2.823	2.762	2.808	2.659	2.764	2.648	2.596	2.368	2.421

Norms

Q	—	1.92	—	—	—	—	—	20.94	27.66	44.04		
or	12.79	8.34	6.12	13.34	14.46	14.46	33.36	15.57	30.02	37.25		
ab	29.34	28.82	29.87	31.96	38.25	33.27	54.49	31.96	33.01	15.20		
an	15.29	25.58	25.02	18.35	15.85	17.51	—	15.85	1.95	—		
ne	—	—	3.41	—	—	5.25	1.99	—	—	—		
C	1.02	—	—	0.41	—	—	—	—	—	0.20		
di	—	3.27	6.80	—	1.36	6.64	1.73	—	0.22	—		
wo	—	—	—	—	—	—	0.35	—	—	—		
hy	14.91	9.76	—	12.82	2.29	—	—	1.85	0.20	0.40		
ol	4.15	—	15.23	4.85	9.74	8.55	—	—	—	—		
mt	6.26	9.98	4.18	6.03	5.34	5.10	0.93	—	—	—		
hm	—	—	—	—	—	—	4.80	7.20	4.32	1.44		
il	6.84	6.08	5.47	5.32	5.17	4.71	0.91	1.67	0.46	0.30		
tn	—	—	—	—	—	—	—	1.57	0.98	—		
ru	—	—	—	—	—	—	—	—	—	0.08		
ap	7.06	2.02	2.35	4.70	4.37	2.69	0.67	2.69	0.17	0.07		
cc	0.20	0.90	—	tr.	tr.	tr.	0.70	0.20	0.50	0.60		
Z	—	—	—	—	—	—	0.18	—	—	—		

Hill and the freshness, the even dense texture, the irregular jointing, and the absence of flow breccia, flow structure, or vesiculation in the basalt suggest that the rock is not a flow but rather a small intrusive.

Petrography.—The Butcherknife basalt is holocrystalline but very fine-grained. The texture is intergranular ophitic with small grains of pyroxene, euhedral crystals of magnetite, alkalic feldspar, and minor analcime interstitial to the lathlike plagioclase crystals. The largest plagioclase crystals measure 0.3 mm. in length, but the average is much less. Magnetite makes up 10–15 per cent of the rock. Apatite also is abundant in small prismatic crystals. The rock is classified as analcime trachybasalt.

Chemical analysis.—The chemical analysis of the Butcherknife basalt (Table 7, No. 3) is characterized by its relatively large content of alkalies. The analysis somewhat resembles that of samples of the Cottonwood Spring basalt and those of the Sheep Canyon basalt (Table 5) and of basaltic rocks from the Terlingua-Solitario region. The Butcherknife basalt closely approaches the composition of an analcime syenogabbro from Cigar Mountain.

Age and correlation.—The Butcherknife basalt is younger than the lowermost Pruett. Although the Butcherknife basalt cannot be correlated definitely with either the Sheep Canyon basalt or the younger Cottonwood Spring basalt, it is much like them in composition.

Igneous Petrology

The chemical analyses of the igneous rocks from the Buck Hill area and the calculated norms (Table 10) bring out the alkalic affinities which characterize the Tertiary igneous rocks of Trans-Pecos Texas. The silica content of the analyzed samples ranges from 44.3 to 78.8 per cent, and total alkalies (Na_2O and K_2O) from 4.8 to 7.5 per cent. Relatively large amounts of TiO_2 and P_2O_5 mark the low-silica lavas. Notable quantities of SrO and BaO were determined in these samples; strontia ranges from 0.09 to 0.14 per cent, and baria, from 0.06 to 0.23 per cent. The Buck Hill rocks are similar in composition to the igneous rocks of the Terlingua-Solitario region described by Lonsdale (1940), although certain rock types from the southern region are not represented in the analyzed specimens from the Buck Hill area, from which new types also are recorded. Most notable of the latter is the Mitchell Mesa rhyolite, a potassic rock in contrast with the sodi-potassic high-silica rocks of the Terlingua-Solitario region. A rough comparison of normative minerals of the two groups follows:

	No. of Analyses	*or*	*ab*	*an*	*ne*	*ol*
Buck Hill	10	6–37	29–55	0–26	0–5	0–15
Terlingua-Solitario	21	8–31	5–55	0–20	0–11	0–16

The analyses from these areas together with three older analyses plotted in the usual silica-oxides variation diagram give relatively smooth curves for the group of 34 samples (Fig. 5). Certain features of the variations are worthy of comment.

(1) Strong curvature of the variation lines suggests strong fractionation of the liquids from which the rocks were derived.

(2) The downward curvature for K_2O in the high-silica range is notable. Points for the curve require little smoothing except for samples of the Potato Hill andesite and the Mitchell Mesa rhyolite. The andesite analysis is anomalous in its abnormally high silica content which is the result of the introduced quartz and chalcedony of the amygdules. By subtracting silica equivalent to the normative quartz and recalculating, the andesite can be plotted at 51.4 per cent SiO_2. The recalculated K_2O then is 3.3 per cent, and this value fits the K_2O curve well. Similarly points for other constituents are shifted to the left largely eliminating the apparent anomalous positions of andesite points plotted on Figure 5. Extrapolation from the main K_2O line to the Mitchell Mesa rhyolite gives a line trending upward in the high-silica range which is in sharp contrast with the plotted curve for the Trans-Pecos rocks but more nearly like the usual variations of K_2O in rock series.

(3) As a group, the analyzed lavas from the Buck Hill area contain relatively more total iron oxide (Fe_2O_3) than the Terlingua-Solitario rocks.

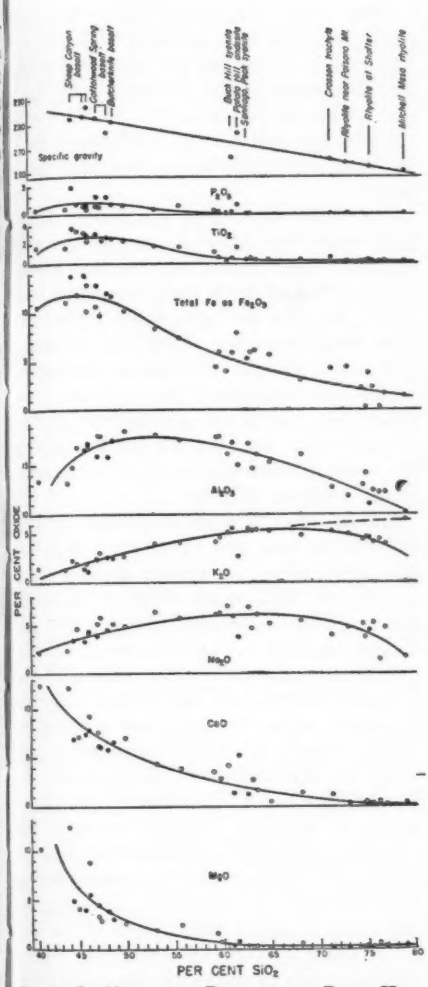

FIGURE 5.—VARIATION DIAGRAM OF BUCK HILL AND SIMILAR IGNEOUS ROCKS

Circles represent analyzed samples from the Terlingua-Solitario region; analyses of rhyolite rocks from Shafter and Paisano Mountain and of syenite from Santiago Peak are included

Broader regional studies of the Cenozoic igneous rocks of Trans-Pecos Texas are now in progress, and a larger number of analyses may show greater diversity and relations other than those suggested by the present analyses. Lonsdale (1940) has discussed the chemical characteristics and origin of the Terlingua-Solitario region comparing them to alkalic subprovinces at the front of the Rocky Mountains.

STRUCTURAL GEOLOGY

General Features

The effects of two crustal disturbances can be seen in the Buck Hill region. The Laramide revolution at the close of the Cretaceous resulted in board upwarping and erosion prior to the deposition of the Tertiary Pruett formation. In the Buck Hill and Elephant Mountain areas there is a discordance between the dips in the Cretaceous which normally do not exceed 4° and in the overlying basal Pruett conglomerate and sandstone. Along the 02 ranch road south of Crossen Mesa dips in the Pruett are as high as 14°, but the beds flatten rapidly northward. Dips in the underlying Boquillas limestone are 2°–3° N. South of the road at this locality the basal conglomerate occupies a swale or depression on the old Boquillas surface. At Hill 3927 northeast of Butcherknife Hill the basal Pruett beds dip 8° N. The high dips in these beds are intrepreted as initial dips off hills on the Boquillas erosion surface.

Faults of late Tertiary or early Quaternary age are the outstanding structural features, but the Buck Hill volcanic series has been deformed into broad open folds. The faults are all of the normal type and cut the youngest rock of the volcanic series and also the syenite intrusives. In the Buck Hill quadrangle there are two major faults or fault zones along which a central block is depressed structurally. The eastern part of the quadrangle is structurally high. Cretaceous rocks are not exposed west of Whirlwind Spring in the northern part of the area and west of The Ditch in the southern part, except at Straddlebug where the Cretaceous strata are domed around the syenite plug. From Whirlwind Spring northwest in the quadrangle and west in the Jordan Gap quadrangle progressively younger rocks of the Buck Hill volcanic series are exposed.

Effects of older orogenies that produced the structure in the Paleozoic rocks in the Marathon Basin (King, 1937) to the northeast and in the Solitario (Sellards, 1932, p. 55–131;

Baker, 1934, p. 177–180) to the south cannot be seen as rocks older than the Georgetown limestone do not crop out in the Buck Hill quadrangle. Likewise, little information concerning the Laramide deformation is available from work in the Buck Hill area; therefore, only the Cenozoic structural features are here considered.

Cenozoic Folding

The Buck Hill volcanic series in the Buck Hill and Jordan Gap quadrangles has been folded into a broad northwest-plunging nose. In the northern part of the Buck Hill quadrangle the regional dip is less than 2° N. or NW. The scarps of the Pruett tuff beneath the Cottonwood Spring basalt and of the Duff tuff beneath the Mitchell Mesa rhyolite face southward toward Green Valley. In the western part of the quadrangle remnants of the Cottonwood Spring flows cap hills north and south of the O2 ranch house, and west of these hills in the Jordan Gap quadrangle the scarps of the basalt-capped cuestas face southeastward toward Green Valley with a gradual shift from an east-west strike in the Buck Hill quadrangle to a north-south strike in the Jordan Gap quadrangle. Similarly, the outcrop of the Mitchell Mesa rhyolite swings southward in the high mesas of the Jordan Gap quadrangle and describes a rough arc with a change from a northerly dip to a westerly and southwesterly dip in western and southwestern Jordan Gap quadrangle. These structural features indicate a broad fold whose axis plunges northwest in a direction roughly parallel to the northwest-trending faults of the western part of the Buck Hill quadrangle.

Cenozoic Faulting

Walnut Draw and related faults.—The Walnut Draw fault zone, named for Walnut Draw on the western edge of Crossen Mesa, consists of roughly parallel and imbricate faults trending southeast that set Crossen Mesa off from the structurally depressed block to the southwest. The fault zone can be traced continuously to the vicinity of Whirlwind Spring, but south of Dark Canyon and east of Whirlwind Spring the master fault is covered by alluvium. From Whirlwind Spring the fault continues to the east boundary of the quadrangle and is marked by the southward-facing Boquillas cuesta. The Walnut Draw fault zone within the quadrange is 15 miles long but continues eastward in the Santiago Peak quadrangle and northward in the Alpine quadrangle.

The displacement along the Walnut Draw fault ranges from 300 feet at the eastern end, where the upper beds of the Boquillas formation are dropped down against the lower part of the Boquillas and against the Buda limestone, to approximately 1000 feet at the northwestern end where the Duff tuff is downthrown against the Cottonwood Spring basalt. Between these points the displacement on any one fault of the zone is small, but the cumulative result is a stratigraphic throw of approximately 1100 feet (Fig. 6).

Walnut Draw is a graben formed by two faults of the zone that are as much as half a mile apart. The downthrown block is tilted and dips 9° NW., so the throw of the faults increases in that direction. Flows above the Crossen trachyte on the mesa occur 400–500 feet lower in the graben, and additional displacement up to several hundred feet is effected by a third fault that makes a horst of the southwestern block of the Walnut Draw graben and brings the Duff tuff down against the Cottonwood Spring basalt. The latter fault is the master of the zone. Vesicular basalt on the southwest side of this fault south of the trachyte-capped spur is mapped as Cottonwood Spring basalt but may be Sheep Canyon basalt, indicating a minimum displacement of 500 feet with a probable maximum of 850 feet. In the fault zone west of the Pruett ranch house the displacement is taken up in a number of faults, and slivers of the Crossen trachyte have a pronounced roll or dip to the southwest indicating that displacement in part was accomplished by bending of the volcanics.

East of Whirlwind Spring the Boquillas limestone forms a southward-facing cuesta in the upthrown scarp along the eastward continuation of the Walnut Draw fault zone. Most of the Pruett formation has been displaced along the concealed fault between Whirlwind Spring and the Boquillas outcrop to the east. This fault, covered by alluvium

between the O2 ranch road and Dark Canyon, is the master fault. It probably parallels the mapped fault on the west side of the Whirlwind Spring spur, but the mapped fault is a minor one, and east of Hill 4631 the Cottonwood Spring basalt is downthrown against the upper breccia-conglomerate member of the Pruett with a displacement of 60 feet.

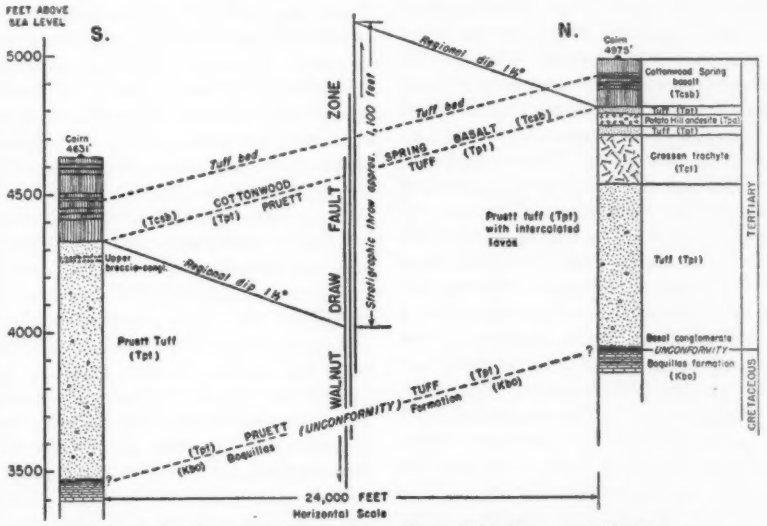

FIGURE 6.—COLUMNAR SECTIONS AT HILLS 4631 FEET AND 4975 FEET
Showing correlation of the Cottonwood Spring basalt flows and interpretation of structure.

Faults of the central area.—South of Mitchell Mesa numerous northwest-southeast faults cut the Cottonwood Spring basalt. The displacement on these faults usually is less than 100 feet, but their net effect is an appreciable lowering of the volcanics to the southwest. The faults cannot be traced to the southeast because of the alluvium in Green Valley, but in the eastern and central parts of the quadrangle three faults with a northwest trend are mapped, and these also result in a structural lowering to the southwest. The large fault southwest of Butcherknife Hill swings eastward and parallels the Chalk Draw fault, forming a graben. Chalk Draw road follows this alluvium-covered structural depression.

Chalk Draw and related faults.—The east-west Chalk Draw fault structurally dominates the southern part of the quadrangle and has a profound influence on the physiography. Although the fault is traced with difficulty in the incompetent beds of the Pruett tuff and in part is covered by alluvium, it extends practically the breadth of the quadrangle and continues eastward in the Santiago Peak quadrangle to Chalk Draw where it swings to the southeast. The fault was named by C. L. Baker (1934, p. 209), and its upthrown scarp forms the west wall of Chalk Draw valley from the southern part of the Santiago Peak quadrangle to the Rosillos Mountains (Fig. 2).

Although the southern upthrown block of the Chalk Draw fault in the Buck Hill quadrangle is structurally high, there is a structural lowering from east to west accomplished mainly by a set of northwest-southeast faults. These secondary faults divide the uplifted block into three segments. The easternmost segment is east of the Alpine-Terlingua highway and is capped by the lower flaggy beds of the Boquillas formation which also form the downthrown block on the north side of the Chalk Draw fault. Northeast of Stone Tank the downthrown block dips at an angle of about 7° toward the fault plane and abuts against the Georgetown limestone. The throw on the Chalk Draw fault

here is approximately 225 feet, but westward the stratigraphic throw decreases, and the Boquillas beds rest against the Grayson and Buda formations.

The middle segment is largely west of the highway and is capped by the upper limestone beds of the Boquillas. Small grabens result from the northwest-southeast faults in this segment; in one a block of the syenite at Buck Hill is downfaulted, and in a second the basal beds of the Pruett tuff lie in a valley flanked by the Boquillas limestone. The downthrown side of the Chalk Draw fault in this segment is the Pruett tuff, and along the fault the tuff beds are characterized by dips up to 40° NW. The high dips suggest that the throw along Chalk Draw fault may be great, although displacement on the fault cannot be determined.

The third segment in the southwest corner of the quadrangle is composed of the Pruett tuff which is downfaulted against the Boquillas limestone south of Loy Place and also is exposed in The Ditch and in Terlingua Creek. In this area the Chalk Draw fault cuts the north slope of Straddlebug Mountain and is exposed to the west in Terlingua Creek where the tuff beds on the downthrown side of the fault dip 36° NW. The high dips in the Cretaceous strata flanking the syenite plug of Straddlebug Mountain are the result of the intrusive action.

The structural lowering from east to west along the Chalk Draw fault is in keeping with the increasing magnitude of structural depression from east to west along the Walnut Draw fault and a similar lowering from northeast to southwest on the faults in the central part of the quadrangle. The northwest dip of the Pruett tuff beds along the Chalk Draw fault may be related to this increasing displacement.

Elephant Mountain faults.—Two parallel faults trending northwest lower the Sheep Canyon basalt and Pruett tuff in the southern tip of the mountain. The displacement on the larger northeastern fault is at least 550 feet. In this corner of the quadrangle the Cretaceous formations attain their greatest altitude, and the top of the Cretaceous is 4550 feet above sea level as compared to 4000 feet above sea level at Buck Hill.

Physiography

The Buck Hill quadrangle lies in the Mexican Highlands Province (King, 1937, p. 2) and affords a marked contrast between its northern part containing residual mesas of the southern Davis Mountain front and its southern part occupied largely by the northeastern end of an extensive lowlands, Green Valley. East of the quadrangle is the folded Santiago Mountain range which is continued to the southeast in the Sierra del Carmen (Fig. 2). North and south of the quadrangle the Alpine and Terlingua areas are important centers of Tertiary intrusive action, and in both these areas the prominent physiographic features are the peaks which are igneous plugs uncovered by erosion.

The semiarid climate of Trans-Pecos Texas brings out the resistant quality of the limestones and igneous rocks which form the prominent topographic features. The mesas in the northern part of the Buck Hill quadrangle are of unequal height and altitude, because different flows cap them and because faulting has disturbed the normal relations of the volcanic series. The mesa tops are essentially dip slopes on flows. Mitchell Mesa capped by rhyolite porphyry is the highest point in the quadrangle with an altitude of 5351 feet, and between this mesa and Green Valley there is a maximum relief of 1700 feet.

Crossen Mesa capped by trachyte porphyry is several hundred feet lower than Mitchell Mesa. Hills rising above the trachyte caprock are capped by flows of the Cottonwood Spring basalt. Potato Hill is of this type, and is the highest point on Crossen Mesa (B. M. 5071 feet). The lower mesas southwest of Crossen Mesa are also capped by the Cottonwood Spring basalt and share the low regional northwest dip. The highest points near their southern scarps generally are less than 4500 feet above sea level. The breccia-conglomerate, 60 feet below the top of the Pruett, makes a prominent bench in the tuff below the Cottonwood Spring basalt. The breccia-conglomerate caps a number of spurs; a prominent one is Boat Mountain (Pl. 3, fig. 2).

Small consequent streams on the mesas flow northward on the dip slopes, but the master drainage is southward and opposed to the regional dip. The northward-flowing streams

on Mitchell Mesa are captured in the Alpine quadrangle by Walnut Draw and Goat Creek and diverted to the Terlingua drainage by way of Duff Creek and The Ditch. South of Duff Springs a trellis drainage pattern controlled by the northwest-southeast faults has been developed, and the tributaries to Duff Creek occupy small grabens. Two miles south of Duff Spring the valley is half a mile wide, but the present stream flows in a narrow channel 30 feet deep cut in the alluvium.

The central and southern parts of the quadrangle consist of rolling hills and cuestas carved in Cretaceous limestone and shale and of valley flats which are alluvium-filled, structurally low areas. The cuestas are fault scarps with the downthrown side commonly covered by alluvium. The most conspicuous scarp is formed by the Georgetown limestone along the Chalk Draw fault. The Boquillas cuesta along the Walnut Draw fault east of Whirlwind Spring has been breached by Calamity Creek. North of the fault scarp on either side of Calamity Creek are low rounded hills which are typical of the Boquillas formation (Pl. 4, fig. 2). Calamity Creek, like Duff Creek, has cut a steep-walled channel in the alluvium of Green Valley. Similar channels characterize much of the courses of The Ditch and Terlingua Creek in the southwestern part of the quadrangle, and these streams are now actively eroding the Pruett tuff bedrock.

The active downcutting with renewal of erosion of bedrock in this region is attributed by Baker (1934, p. 140) to recent rejuvenation. According to Baker the development of the Rio Grande involved a number of local base levels which have left their record in "erosion-cut or 'pediment' surfaces of the bed rocks in numerous places". Remnants of terraces or benches approximately 50 feet above the present valley level can be seen west of the Alpine highway in the northern part of the quadrangle and in a number of places in the southwestern part. These benches are erosion-cut surfaces on well-indurated Pruett tuff beds with a thin veneer of gravel.

Summary Geologic History

The Cretaceous record in the Buck Hill quadrangle is limited; it ranges from Georgetown limestone to calcareous shale of Eagle Ford or lower Austin age. South of the quadrangle, Cretaceous formations (Adkins, 1932, p. 451, 505–512) younger than the Boquillas include the Terlingua, Aguja, and Tornillo formations. The Terlingua and the lower part of the Aguja are marine, but the upper Aguja beds contain black carbonaceous and lignitic shales that mark a change from marine to brackish-water and nonmarine deposition. Remnants of the Terlingua and Aguja formations occur south and east of the Buck Hill quadrangle. Calcareous shale at Hill 3927 northeast of Butcherknife Hill contains lower Austin Foraminifera and may be a remnant of the Terlingua (Austin) formation.

Uplift in late Cretaceous time initiating the Laramide revolution resulted in erosion of the Terlingua, Aguja, and the Tornillo if the latter was deposited in the Buck Hill area. The basal Pruett conglomerate, which rests on the Boquillas limestone in the Buck Hill quadrangle, lies on the Terlingua clay in a number of places in the Agua Fria quadrangle and on the Aguja formation on either side of the Alpine-Terlingua highway southeast of Hen Egg Mountain in the northern part of the Terlingua quadrangle, approximately 20 miles to the south. North of the quadrangle in the vicinity of Mount Ord, the volcanics rest with "well-marked unconformity" (King, 1930, p. 98) on rocks of Washita age. These observations indicate the widespread angular unconformity between the Cretaceous strata and the Tertiary volcanics, with the volcanic series resting on progressively younger Cretaceous rocks as one goes south toward the Chisos Mountains. The persistence of the basal Tertiary conglomerate suggests that a similar break might be expected between the Tornillo formation, if it is of Cretaceous age, and the Chisos beds in the Chisos Mountain area.

West of the Buck Hill region in western Presidio County (Fig. 2) the effects of the Laramide revolution are more pronounced, and the Tertiary volcanics rest on Permian strata in the Shafter area and on folded Upper Cretaceous formations in the Rim Rock country of the Sierra Tierra Vieja. In the latter area near Provenir vertebrate fossils date the Vieja volcanic series as lowermost Oligocene. Abun-

dant igneous boulders in the basal Vieja conglomerate indicate Eocene or possibly late Cretaceous igneous activity with erosion of intrusives and possibly flows to supply the boulders. In contrast with the Vieja conglomerate the basal Pruett conglomerate in the Buck Hill region is composed of well-rounded pebbles and cobbles of limestone and chert. Large angular pieces were derived from the formations immediately below the conglomerate, but igneous pebbles are few or wanting. The first appearance of large igneous pebbles and boulders is in the upper breccia-conglomerate in the upper part of the Pruett tuff, and at approximately this level the intercalated flows appear in the sequence.

The Pruett is assigned tentatively to the Eocene on the basis of fossil gastropods in the fresh-water limestone beds that occur below and above the Crossen trachyte and between flows of the Sheep Canyon basalt. Therefore, these flows can be dated tentatively as Eocene. The younger Potato Hill andesite also may be of Eocene age, but there is no definite evidence for dating the time of its extrusion or of that of the overlying Cottonwood Spring basalt flows. The Duff formation is assigned tentatively to the Oligocene. It is not likely that deposition was continuous, and undoubtedly materials were reworked and locally were removed by streams. Rather large breaks in the tuffs could escape detection because of the nature of the materials.

The various flows in the Buck Hill volcanic series were all more or less weathered and eroded. The Crossen trachyte in Sheep Canyon (Pl. 6) was deeply incised, and the younger Sheep Canyon basalt flows which filled the ancient stream channels rest on fresh-water limestone and on trachyte. The Potato Hill andesite also was deeply weathered, and the variable thickness of the Cottonwood Spring basalt, in part due to thinning or feathering out of the flows, probably is also the result of erosion prior to deposition of the Duff tuff.

Stratification of the tuff and the fresh-water limestone beds indicate that much of the Pruett was deposited in water, probably in large lakes. The calcium carbonate, especially in the lower part of the formation, may have been mechanically derived from the exposed Cretaceous formations, but the thick sequence of tuffaceous limestone in Sheep Canyon and in the Elephant Mountain area was precipitated. The algal structures and charophytes suggest that much of the limestone is of organic origin. The Tertiary lakes probably were temporary and shallow, and the lavas do not show structures or alteration commonly found in subaqueous flows. The source of the volcanic ash is problematical. The finer material could have come from great distances, but the coarser detritus, such as the breccia-conglomerate in the upper Pruett and the conglomerates of the Duff, probably were derived from near-by sources.

In late Tertiary time syenitic and basaltic magmas intruded the Cretaceous and Tertiary formations. The syenite at Buck Hill intruded the lower Pruett beds, and the Elephant Mountain syenite invaded upper beds, coming to rest on the Sheep Canyon basalt. The syenite of Santiago Peak stands 1200 feet above the highest level of Cenozoic tuff beds that encircle its base. In the southwestern part of the Jordan Gap quadrangle igneous plugs intruded the Duff formation and the younger tuff beds above the Mitchell Mesa rhyolite. These intrusives may not all be of the same age, but they indicate activity in late Tertiary time; however, they antedate the late faulting. In the Marathon basin numerous small igneous intrusives commonly are closely related to the structure, and King (1937, p. 117) points out that the Tertiary intrusives "appear to have come up along the planes of thrust faults of Paleozoic age". The alignment of the syenite intrusives at Buck Hill and Straddlebug Mountain along the Chalk Draw fault is interesting in this respect, and suggests that the fault along which movement took place in late Tertiary time was an older zone of weakness along which the syenite intrusives were emplaced.

The time of folding and faulting of the Buck Hill volcanic series cannot be demonstrated from the data at hand. It is at least late Tertiary and older than the basin fill and the alluvial deposits of the region. The structural history of Trans-Pecos Texas is complex, and a full understanding will be gained only with solution of some of the larger structural prob-

lems which have been summarized by Baker (1934; 1941) and King (1935; 1937).

Since the late Tertiary uplift, erosion has removed a great thickness of the volcanic series, forming Green Valley and again exposing the Cretaceous strata. Terraces and alluvial deposits indicate that the process has been controlled by temporary base levels, and the present active downcutting indicates recent rejuvenation.

Economic Geology

There are no ore deposits in the Buck Hill quadrangle, and there is little local demand for rock aggregate or building stone. A number of tests drilled in exploration for oil and gas in the region have been unsuccessful. Water supply is a local problem in part taken care of by construction of tanks to store the surface run-off for watering cattle. The tanks are earthen or more rarely concrete dams thrown across drainage ways with the storage capacity commonly increased by scooping out a depression. Water also is obtained from wells equipped with windmills. In the northwestern part of the quadrangle the springs which issue from the flows of the Cottonwood Spring basalt have been the main source of water which is piped to the southern part of the area. The water in the Cottonwood Spring basalt has been developed recently by the O2 ranch, and a number of wells up to 300 feet in depth were completed in the area of the basalt south of Mitchell and Crossen mesas. A good well at the Pruett ranch house obtains water from the Walnut Draw fault zone. Water at the Kokernot ranch is drawn from wells in the alluvial gravel.

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Appendix

Detailed Measured Sections

Section No. 1

Section of the Buda limestone and the upper part of the Grayson clay

Measured in the west bank of Calamity Creek at a point 4½ miles south of the Kokernot ranch house.

	Thickness Feet	Inches
Buda formation:		
17. Limestone, gray, porcelaneous, thick-bedded, brecciated; weathering to a dark-gray etched surface	25	
16. Limestone and marl; limestone, gray, nodular, thin-bedded, fine-grained, fossiliferous; marl, buff to gray, weathering to light tan, fossiliferous. Ammonites, echinoids, etc.	20	
15. Limestone, gray, thick-bedded, brecciated; lower surface undulating	12	
Grayson formation:		
14. Clay, blue-green, brown, red; oxidized along joints and cracks; some thin sandstone layers	2	
13. Clay, green to buff; variegated green, blue, buff, and yellow on weathering; compact to laminated, grading to shale	2	
12. Limestone flags and marl, similar to unit No. 9	5	
11. Marl and clay; weathers to tawny yellow	2	
10. Limestone and marl; limestone gray, weathering white, nodular, contains *Haplostiche texana;* clayey beds tawny yellow with abundant echinoids and *Exogyra*	8–12	
9. Limestone and marl; limestone flags, nodular, lenticular, dark gray-blue, sandy weathering to reddish sandy surfaces; marl, sandy and clayey, buff, weathering to light tawny-yellow to reddish buff; erosion leaves ledges of the limestone flags	2	6
8. Marl, sandy, buff, fine-grained, soft, some thin limestone lentils		3-6
7. Clay, green to buff, compact or slightly shaly		6
6. Shell bed; composed almost entirely of *Haplostiche texana* with sandy matrix, weathering to buff and tawny yellow		1-2
5. Sand, shaly, weathers to bright yellow, dull blue and buff; fine-grained, friable		2
4. Clay, light-buff, greenish-yellow, weathering to tawny yellow and variegated colors, forms receding slope	2	2
3. Limestone, flaggy, lenticular, dark gray-blue, medium-grained, sandy, weathering reddish buff		2
2. Shale, light greenish-yellow to olive-green, weathering to very light-buff, grading laterally to sandy limestone flags		3-4
1. Limestone and clay	2	
Total thickness measured	77	

Lower part of Grayson covered.

Section No. 2

Section of the upper part of the Grayson clay in east bank of large fault gully at southeastern foot of Elephant Mountain

	Thickness Feet	Inches
Buda formation:		
13. Limestone, gray, fine-grained, nodular; lower surface uneven, undulating	3	

	Thickness Feet	Thickness Inches
Grayson formation:		
12. Clay, and shale; alternating brown shale and white clayey material, laminated; weathers dull tannish white		6–8
11. Sandstone, fine-grained, stained reddish		2
10. Shale, tan to brown, weathering to lighter color, laminated		5
9. Sandstone and shale alternating, reddish sand in layers less than 1 inch thick; brownish-red shale	1	3
8. Sandstone and shale with abundant fossils: *Exogyra cartledgei*, *Exogyra whitneyi*, *Ostrea* sp., *Pecten* sp. Shells commonly broken; sand, fine, tan to reddish brown	1	
7. Shale, sandy, brown, fine-grained, laminated	1	3
6. Limestone, gray, nodular, even-textured, slightly sandy near top		6–8
5. Shale and sandstone; thin-bedded, brown, fine-grained, weathers tan, friable		7–9
4. Shell breccia with calcium carbonate cement		2–3
3. Shale, some sand, brown, laminated		8
2. Sandstone, calcareous, fine-grained, well cemented		2
1. Limestone and shale; alternating layers; nodular limestone, very fossiliferous near top with *Inoceramus*, echinoids, and gastropods; weathers to tan and dark buff	4	
Total thickness measured	14	

Section No. 3

Section of the Boquillas formation

Measured in gully on south side of Elephant Mountain 3 miles northeast of the Kokernot ranch house, beginning at small fault just north of water tank.

	Thickness Feet	Thickness Inches
Boquillas Formation:		
39. Covered interval, may be Boquillas beneath Pruett tuff	20(?)	
38. Limestone, bluish-gray, fine-grained, a little marl, weathers chalky white	5	
37. Limestone, light-gray to buff, fine-grained, in 1-foot beds separated by marly layers not over 4 inches thick	21	
36. Limestone, light-buff, fine-grained, weathers buff	2	
35. Limestone, fine-grained, slightly nodular beds approximately 1 foot thick separated by thin marly layers, contains *Inoceramus*	8	6
34. Limestone, thick bed, gray to light-buff or tan, very fine-grained, contains small limonite particles, forms overhanging ledge, contains *Inoceramus* fragments	2	6
33. Limestone and marl alternating in 4-inch beds, limestone slightly nodular	2	
32. Limestone, 1-foot beds separated by thin marl layers	5	6
31. Limestone, light-gray with buff tinge, fine-grained, weathers buff to light yellow, contains impressions of *Inoceramus*	2	6
30. Limestone, and marl interbedded; limestone as in interval above, averaging 8–10 inches thick; marl slightly yellow, gray near top	3	
29. Limestone, light-gray to buff, fine-grained	1	
28. Limestone and marly shale, alternating in beds up to 6 inches thick; limestone, gray; marl, buff; *Inoceramus* casts	6	
27. Limestone and marl interbedded in layers averaging 2 inches thick; marl beds more uniform in thickness; colors as above		5
26. Limestone flags and shale alternating, contacts undulating with thickening and thinning, shale weathers to platy structure; flags 2 inches thick, except for 4-inch flag topping interval; banding results from weathering of flags. Many *Inoceramus* casts	8	6
25. Flags and marly shale, alternating as above with 1-foot limestone layer at top of interval, ½-inch layer of *Inoceramus* shell fragments beneath	8	
24. Marly shale with interbedded limestone flags	1	6
23. Flags and shale alternating, bluish-gray flagstones averaging ½ inch thick; shale has pronounced limonitic stains on bedding planes, average 3 inches thick; fossil ice markings on upper shale layers; flags weather dark gray, shales buff	2	

	Thickness Feet	Inches
22. Marly shale and flags; flags average 1 inch thick, marly layers 3–4 inches; gray to bluish-gray flags weather gray to buff, light-gray marl weathers buff	7	
21. Marly shale and limestone flags alternating, flags ½ inch to 1½ inches thick, gray to dark blue-black weathering gray to buff; marly layers up to 2 inches thick, rather hard, fine-grained, even-textured, weathering light tan and gray	3	
20. Limestone flag, light bluish-gray, very fine-grained, weathering light tan to buff, forms overhanging ledge		4
19. Flags and shaly marl alternating, slightly nodular flags 1–2 inches thick, marl 1–3 inches thick, weathering buff with platy structures	4	
18. Limestone flag, bluish-gray with ½-inch dark blue and green band near top		4
17. Limestone flags averaging 2 inches thick alternating with laminated marl in beds 8–10 inches thick, weathering light tan to yellow with platy structure	8	
16. Limestone flag, light-gray weathering light grayish brown, in places laminated or banded		4
15. Marl and shale with marl predominating, light-gray to white, laminated to thinly bedded	3	8
14. Limestone flag, light-gray, very fine-grained, weathers to dark gray		5
13. Covered, probably thin-bedded limestone and shale	3	6
12. Limestone flag, blue-gray, very fine-grained, sandy; weathers to sandy surface and banded appearance		4
11. Shale and shaly marl, gray to light bluish-gray; weathers with platy structure to light gray or white	3	
10. Flags alternating with shale and thinner limestone layers. Flagstones average about 4 inches thick, shale with limestone layers about 2 inches thick. Shale contains abundant *Inoceramus* prism fragments, commonly stained with limonite	2	6
9. Limestone flag, dark-gray to light-brown, sandy		2
8. Limestone flag, grayish-blue, very fine-grained with little fine sand, some banding, weathers to sandy surface		5–7
7. Shale and thin sandy limestone layers alternating irregularly, very thin-bedded		6
6. Limestone flag, gray, very sandy, in places banding brought out by stains of limonite		4
5. Shale, laminated, light-tan weathering grayish white; minor gray to blue sandy limestone		4
4. Limestone flag, gray, banded, very sandy		6
3. Shale, laminated, tan, sandy		4
2. Limestone flag, gray, banded, sandy, weathers dark gray with rough sandy surface		6
Buda formation:		
1. Limestone, gray, massive, weathers with rough etched surface, very fine-grained, porcelaneous	38	
Total thickness measured	155	

Section No. 4

Partial section of the Pruett formation below the Crossen trachyte

Measured in first gully north of the southeastern end of the Crossen Mesa 2 miles southwest of the Kokernot ranch house.

	Thickness Feet
Pruett formation:	
15. *Crossen trachyte intercalation:* Dark reddish-brown flow of trachyte porphyry with conspicuous columnar jointing	115
14. Covered, in part fresh-water limestone as indicated by float	20
13. Tuff, fine-grained, dove-colored, banded	4
12. Tuff, white, fine-grained, forms receding slope	32
11. Tuff, coarse gray to white beds in less indurated finer-grained ash	81
10. Tuff, gray, massive, forming an overhanging bluff	30
9. Tuff, fine-grained, nodular, hard, blue-gray	146
8. Covered in line of section, tuff	76
7. Tuff, massive, weathering to a rough nodular gravel	24

	Thickness Feet	Thickness Inches
6. Covered in line of section, tuff		20
5. Tuff, massive, gray to buff with large amount of white ash fragments, hard, weathers to a porous surface		18
4. Tuff, gray, fine-grained with small flakes of biotite, massive, weathers to a nodular hard surface, forms an overhanging cliff		26
3. Tuff, extremely hard, dark grayish-blue, dense, weathers to bluish gray		2
2. Tuff, similar to unit No. 4		9
1. Tuff, gray to white		10
Total thickness measured		613

Base of section covered by alluvium, approximate altitude, 4200 feet.

Section No. 5

Section of the Cottonwood Spring basalt flows and of the upper beds of the Pruett formation in vicinity of elevation 4054 feet south of Cottonwood Spring

	Thickness Feet
Cottonwood Spring basalt:	
16. Basalt flow No. 7; thick, massive basalt capping mesa and forming high cliff in southern scarp. Upper amygdaloidal part 20 feet thick grades down to coarse-grained basalt. Fresh color reddish gray weathering to gray and yellowish brown. Lower part of flow forms slope with only scattered outcrops and with abundant ocotillo. Vesicular flow breccia at base	164
15. Tuff, light-gray to red; red rock is hard, dense, baked tuff with subconchoidal fracture. Basalt fragments roughly rounded in lower 1–2 feet of gray to pink tuff. Basalt is vesicular and in pieces up to 4 inches in diameter	8–19
14. Basalt flow No. 6; massive, weathering with slabby structure in upper part and spheroidally in lower part; upper part is 14-foot thick amygdaloid	46–51
13. Basalt flow No. 5; upper 6 feet amygdaloidal with fillings of calcite and chalcedony; lower part massive, uniform	10–12
12. Basalt flow No. 4; amygdaloidal top about 4 feet thick, reddish-gray; lower part, dense basalt	12–13
11. Basalt flow No. 3; amygdaloidal top 2½ feet thick with abundant fillings of chalcedony and of agate with small quartz crystals in cavities at centers of fillings. Base of flow is	

	Thickness Feet
flow breccia, 2 feet thick; middle part, dense and uniform	6–9
10. Basalt flow No. 2; vesicular top with amygdules of chalcedony, calcite, abundant red and yellow jasper.	6–15
9. Basalt flow No. 1; basal flow of the Cottonwood Spring basalt; upper 9 feet amygdaloidal reddish-brown, ropy (pahoehoe), with many fillings of calcite, chalcedony, and agate. Green chloritelike coating on basalt in cavities and around amygdules; fillings of agate commonly terminated inward with small quartz crystals. Lower part of flow is massive and forms prominent bench or ledge 3–4 feet thick of coarse-grained, black basalt with plagioclase needles aligned in flow structure. Flow is badly weathered, but fresh samples are available from heavy ledge and from large boulders. Analyzed specimen No. 840, Table 7, No. 2	42–50
Pruett formation:	
8. Tuff, gray to reddish-brown with conspicuous light-gray to white mottling, even earthy fracture, baked to hard, reddish-brown rock with subconchoidal fracture immediately below basalt flow; contact with overlying basalt sharp and distinct	48
7. Tuff, gray to pink, thin-bedded with upper beds somewhat thicker than lower ones, indurated	16
6b. *Breccia-conglomerate:* massive ledge, 5½–6 feet thick, corresponds to 6a below. Section was offset at this point. Beds from 6a and down were measured below elevation 4054; beds from 6b and up were offset approximately 400 feet to the west	
6a. Massive breccia and conglomerate of igneous rock pebbles and fragments capping spur at elevation 4054′ south of Cottonwood Spring and continuing eastward to Whirlwind Spring and westward into the Jordan Gap quadrangle. Fragments generally 0.1–0.2 inch across but many larger up to boulders of 1 foot or more in diameter. Chiefly igneous rock fragments with some chert and banded red and yellow jasper in a gray to yellow matrix of volcanic ash, locally calcareous. Weathers to a porous nodular, yellow to dark-brown ledge commonly covered with ocotillo. Varies laterally both in thickness and in size of component fragments, but is a prominent ledge maker and horizon marker	2
5. Volcanic ash, gray to light-yellow, roughly bedded, and essentially one	

	Thickness Feet
massive layer; fine-grained, with rough uneven fracture, weathers to dark yellow brown	13
4. Tuff, purplish-gray, hard, brittle, banded	1
3. Ash, pink to reddish, mottled with white, weathers to a smooth slope	23
2. Ash, gray to red, poorly bedded, in part well lithified. Beds from 6 inches to 1 foot thick. Alternation of indurated and softer beds produce a step-like appearance in outcrop	7
1. Ash, gray to red, weathers to smooth slope usually covered with débris	16
Total thickness measured	437

Thickness of the Cottonwood Spring basalt, 300± feet.
Base of section, 3992 feet above sea level.

Section No. 6

Section of the Sheep Canyon basalt flows

Measured by G. C. Hardin, Jr., in Sheep Canyon, Alpine quadrangle.

	Thickness Feet
Cottonwood Spring basalt:	
12. Basalt flow, amygdaloidal at top, oxidized and badly weathered	33
11. Basalt flow, amygdaloidal at top, thin flow breccia at base	98
Pruett formation:	
10. Tuff, pink at base and blood red near top	3
9. *Potato Hill andesite intercalation:* Massive porphyritic andesite near base; upper 30–40 feet flow breccia	72
8. *Sheep Canyon basalt intercalations:* Porphyritic basalt flow No. 4	28
7. Tuff, light-pink grading to light green	4.5
6. Porphyritic basalt flow No. 3; amygdaloidal near top; phenocrysts locally segregated (glomeroporphyritic)	42.5
5. Fresh-water limestone, light-gray to white, tuffaceous	6
4. Basalt flow No. 2; coarse even-grained with a few phenocrysts, amygdaloidal near top	32
3. Tuff with very thin lenticular beds of fresh-water limestone	5.5
2. Basalt flow No. 1; dense, even-grained, amygdaloidal near top. Possibly more than one flow as section is partially covered with talus	131
1. *Fresh-water limestone:* dense, gray, massive, tuffaceous limestone	
Total thickness measured	455.5

Section No. 7

Section at Hill 4975 on Crossen Mesa

	Thickness Feet
Cottonwood Spring basalt flows:	
11. Basalt flow No. 6; dense, black basalt with prominent flow breccia at base	55
10. Tuff, red, dense, hard, baked	7
9. Basalt flow No. 5, amygdaloidal at top	7
8. Basalt flow No. 4, amygdaloidal at top grading down to coarse-grained basalt, weathers to yellowish brown	16
7. Basalt flow No. 3, amygdaloidal at top with blowholes ½ inch or more across, fillings of calcite and chalcedony	11
6. Basalt flow No. 2, amygdaloidal at top 2–3 feet thick	15
5. Basalt flow No. 1, amygdaloidal at top about 14 feet thick; fine-grained, somewhat vesicular flow breccia at base about 12 feet thick	61
Pruett formation:	
4. Tuff, grayish-white, fine- to medium-grained	26
3. *Potato Hill andesite intercalation:* porphyritic andesite flow, red residual material and weathered vesicular andesite at top, upper part of flow altered green, many amygdules of chalcedony; grades down to reddish-brown vesicular andesite flow breccia which weathers to greenish and yellowish-brown boulders. Massive andesite at base of flow with numerous plagioclase phenocrysts up to 1 inch in length and ½ inch across. Ocotillo common on weathered slopes of andesite. Analyzed specimen No. 812, Table 6	40
2. Tuff, light yellowish-gray to greenish-gray with numerous light-gray fragments. Analyzed specimen No. 813, Table 2, No. 1	37
1. *Crossen trachyte intercalation:* porphyritic trachyte flow with many stubby feldspar phenocrysts in a dense red groundmass	
Total thickness measured	275

Base of measured section, 4699 feet above sea level.

Section No. 8

Section of the Duff tuff measured in largest gully on south slope of Mitchell Mesa

	Thickness Feet
Mitchell Mesa rhyolite:	
24. Rhyolite porphyry, phenocrysts of quartz and glassy sanidine in dense to vesicular grayish-pink groundmass	38
Duff formation:	
23. Covered in line of section, tuff	64
22. Tuff, white, massive, fine-grained	35
21. Tuff, gray to buff and light-brown, roughly bedded and cross-bedded, medium-grained	104
20. Conglomerate, well-rounded pebbles in matrix of coarse tuffaceous sand	8–10
19. Tuff, gray to light-buff, cross-bedded, fine-grained	46
18. Conglomerate, gray, roughly bedded and cross-bedded; pebbles, cobbles, and boulders up to 1 foot in diameter, mainly of igneous volcanic rocks ranging from rhyolite to basalt, some of vesicular lava, commonly well rounded. In places cobbles about 6 inches in diameter are well sorted with little or no matrix material	6–8
17. Tuff, light-buff to reddish-brown, locally gray; medium-grained, bedded in layers 4–6 inches thick; stands with vertical front	105
16. Tuff, grayish-white with several 6-inch beds of brownish-red stained ash, fine- to medium-grained, massive	123
15. Tuff, stained light brownish red, with small white spots, coarse- to medium-grained, forms a prominent bench	4–6
14. Tuff, gray with several 6-inch to 1-foot reddish-buff beds near top, fine- to medium-grained, massive and holds a steep face on scarp; weathers to small nodular forms. Many thin clastic dikes of dark-gray to buff-colored sandy ash cut this unit	238
13. Tuff, buff to light-brown, medium-grained	2–4
12. Tuff, gray, fine-grained, massive, cut by many small brown, gray and varicolored clastic dikes	29
11. Tuff, buff, coarse-grained; forms benches; grades laterally to a thin breccia	3
10. Tuff, light- to dark-gray, fine-grained, weathers to a nodular, concretionary surface; stands in steep faces and forms benches	125
9. Tuff and breccia, dark-buff with small white ash flecks; forms a flat surface on weathering	3
8. Tuff, light-gray, fine-grained, uniform, massive, nodular on weathering	17
7. Volcanic breccia, light- to dark-buff, with fragments several millimeters across in fine-grained ash matrix. Forms ledge and overhanging layer.	2–3
6. Tuff, light-gray to tan, in most places peppered with flakes of biotite, weathers to nodular surface, in places cut by clastic dikes	53
5. Conglomerate, generally stained brown, great range in size of component pieces which are largely volcanic igneous rocks, from small pebbles to boulders 6 feet across. Matrix commonly cross-bedded coarse sandstone. In the line of section this conglomerate bed is relatively thin, but in places it is up to 30 feet thick	6
4. Tuff, dark-gray to buff, very coarse, roughly bedded and cross-bedded, grades to sandstone in places	20
3. Tuff, gray, fine-grained, massive, weathers to a porous honeycombed surface. Forms vertical face, commonly undercut	9
2. Covered in line of section, tuff	8
Cottonwood Spring basalt:	
1. Basalt flow, fine-grained, very vesicular at top	5
Total thickness measured	1058

Section No. 9

Section of Sheep Canyon basalt flows and Pruett tuff

In gully west of first large fault northeast of the Kokernot ranch house on the southern slope of Elephant Mountain.

	Thickness Feet
12. Covered interval, may possibly be basalt	50
11. *Sheep Canyon basalt intercalation:* basalt porphyry flow, black, weathering to rusty brown. Vesicular with openings in part filled with calcite, vesicles elongated in direction approximately N. 65° W.	171
10. *Sheep Canyon basalt:* porphyritic flow, upper 20–30 feet vesicular and amygdaloidal with fillings of jasper, chalcedony, and calcite. Abundant glassy, green, plagioclase phenocrysts up to 1 inch in length. Lower	

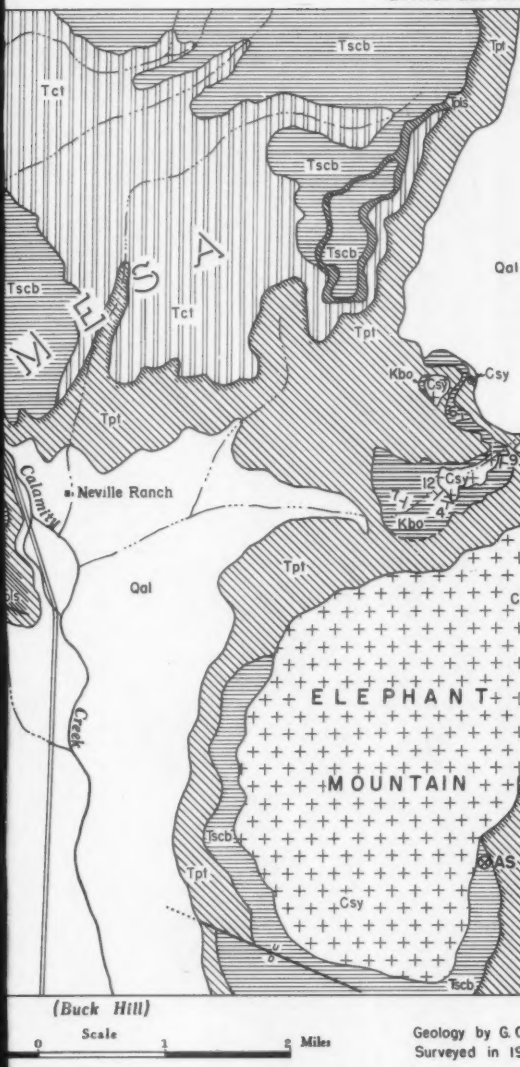
	Thickness Feet
part of flow is massive, weathering spheroidally. Gray, medium-grained tuff occurs between this flow and the upper basalt in some places.	121
9. Fresh-water limestone, light-gray, very fine-grained, in beds 1–2 feet thick. Weathers to chalky white, forms prominent ledges. Gastropods and shell fragments in part silicified. Thickness variable; may be as much as 13 feet	4
8. *Sheep Canyon basalt:* porphyritic flow, vesicular in upper half with many calcite amygdules. Limestone above flow contains pebbles of basalt and extends down filling crevices and openings in upper part of flow	88
7. Covered interval, probably tuff	83
6. Tuff, hard, well-indurated, gray to white, fine-grained, maintains vertical faces	70
5. Tuff, indurated, light-gray, medium-grained	10
4. Tuff, hard, gray to light-yellow	32
3. Conglomerate, chert and hard tuff pebbles	4.5
2. Tuff, hard, gray, uniform	8
1. Tuff, gray to pink or purplish, massive, weathers to nodular, pebbly surface. Some pebbles of dense gray limestone. Molds of low-spired fresh-water snails	84
Total thickness measured	725.5
Base of section covered by alluvium.	

Topographic base by U.S. Geological Survey.
Surveyed in 1893.
Edition of 1895.

Scale 1 0 1 2 Miles

Geology by
Surveyed i

GEOLOGIC RECONNAISSANCE MAP OF THE ELEPH MOUNTAIN AREA, BREWSTER COUNTY, TEXAS



Geology by G. C
Surveyed in 19

SANCE MAP OF THE ELEPHA
BREWSTER COUNTY, TEXAS

dich and Elms, Pl. 6

(Monument Spring)

30°00'

103°30'

(Santiago Peak)

Geology by G. C. Hardin, Jr.
Surveyed in 1941.

LEPHANT
EXAS

EXPLANATION

QUATERNARY

Qal
Alluvium

CENOZOIC

Csy
Intrusive syenite

TERTIARY

Tcsb
Cottonwood Spring basalt

Tpt
Tscb
Tct
Tpa
Tpis
Pruett tuff formation

(Volcanic tuff breccia, and (Tpis) fresh-water limestone member; intercalated flows: (Tpa) Potato Hill Andesite, (Tscb) Sheep Canyon basalt; (Tct) Crossen trachyte)

UNCONFORMITY

UPPER CRETACEOUS

Kbo
Boquillas formation

6
Strike and dip of beds

U
D
Normal Fault
(U, uplifted side; D, downthrown side, line solid where known; dotted where inferred under cover of alluvium)

FL
Fossil locality

Line of measured section
(Appendix, No. 6)

⊗AS
Analyzed sample
(Table 5, No. 3)

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GEOLOGY OF THE MANZANITA AND NORTH MANZANO MOUNTAINS, NEW MEXICO

By Parry Reiche

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Abstract

The Manzanita and North Manzano mountains, segments of a north-south fault-block range in central New Mexico, consist of low-dipping Pennsylvanian marine strata, chiefly limestones, which rest on a peneplaned Proterozoic sequence of slightly metamorphosed clastic sediments and acidic and basic volcanics, and a granitic stock. The fault-block range has developed since middle Tertiary times. Previous disturbances include rather high-angle thrusting of late Precambrian and Laramide (?) age, the former from the south, the latter from the west. A broad piedmont zone west of the mountains bevels Permian and upper Triassic sediments which are cut off to the west along the trace of a Recent fault.

Introduction

The northeasternmost basin range in the United States trends northerly for 80 miles in central New Mexico. It only locally exceeds 4 or 5 miles in width and rises steeply 1200 to 4500 feet from the piedmont bordering the Rio Grande Valley to the west. The eastern flank is less steep and high. The range comprises four segments; from north to south they are the Sandia Mountains (20 miles long), the high, flat-topped North Manzano Mountains (14 miles), the South Manzano Mountains (19 miles), and the less rugged Los Pinos Mountains (15 miles). The first two of these are separated

by the Manzanita Mountains, a 12-mile stretch in which the Sacramento section of the Basin and Range Province (Fenneman, 1931, p. 393–394) occupies the divide. These five features differ sufficiently in structure to justify individual study.

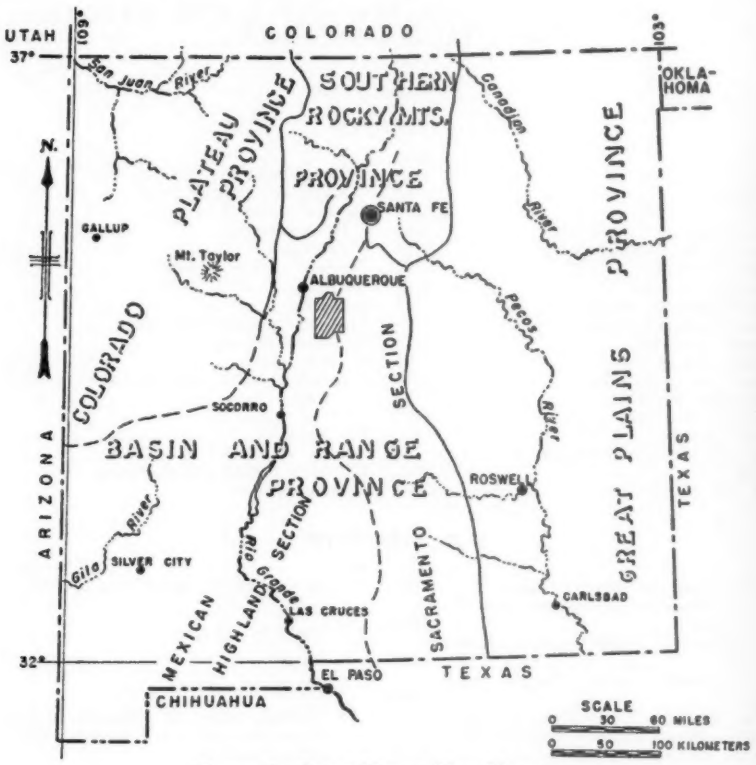

FIGURE 1.—INDEX MAP OF NEW MEXICO
Showing major physiographic divisions (after Fenneman) and area described in this paper (shaded).

Near Albuquerque, New Mexico, the Basin and Range Province narrows rapidly northward and comprises the Sacramento section on the east, the range described, and a complex graben, the Rio Grande Depression, about 25 miles wide, which continues northward beyond the Sandia Mountains into Colorado (Bryan, 1938, p. 198).

The area mapped (Pls. 1, 5) comprises about 250 square miles. The only community is a war-developed ordnance experiment station; less than half a dozen other buildings were occupied throughout 1945. However, the district is readily accessible, and its comparative neglect by geologists is due to the lack, until recently, of available base maps. Lee and Girty (1909) described the district, and Ross (1909) studied the northern half of the Manzanita foothills. Staatz (1912) published a singularly misleading note on conditions immediately south of Hell Canyon. Darton's reconnaissance studies of the State (Darton, 1910; 1922; 1928) include descriptions and maps of the range, and one of his cross sections (1922, p. 219, D–D′) coincides approximately with one of this report (Pl. 5, Sec. C–C′). His geologic map of New Mexico on a scale of 1:500,000 was published in 1928. Ellis (1922) did reconnaissance immediately to the north. More recently, central New Mexico has been aerial-

photographed for the Soil Conservation Service so that contact prints, mosaics, and planimetric sheets have been available. C. B. Read and his coworkers of the U. S. Geological Survey used these and have published maps clarifying Pennsylvanian and Permian stratigraphy over wide areas. One map (scale 1:66,000) includes the area here described (Read, Wilpolt, *et al.*, 1945). Stark and Dapples (1946) have detailed the geology of the Los Pinos Mountains.

This paper utilizes Read's stratigraphic findings and stresses structure and the Precambrian formations. Planimetric control was available except for the area west of Long. 106° 30′; locations in that strip are therefore inferior.

Courtesies were generously extended by the local officials of the Soil Conservation Service, the Forest Service, and the Indian Service; by Mr. Esquepula Jojola of Isleta and by Dr. E. J. Workman and his staff of the Naval Research project, now sponsored by the New Mexico School of Mines at Socorro. The University of New Mexico partially defrayed expenses of drafting and thin sections. Professor Frederick K. Morris of the Massachusetts Institute of Technology has kindly reviewed many critical thin-section descriptions and classifications.

Geographic Features

The North Manzano and Manzanita mountains lie in Bernalillo, Valencia, and Torrance counties, New Mexico, between 35° 02′ 30″ and 34° 42′ 30″ N. Lat. and 106° 20′ and 106° 33′ W. Long. They include the eastern portion of the Isleta Indian Reservation. Much of the higher country is within the Cibola National Forest. Except for the southernmost 5 miles, the eastern boundary of the area follows the watershed divide between the Rio Grande to the west and the closed Estancia Basin. Elevations above sea level average 5400 feet for the west edge of the area, 6000 for the inner margin of the piedmont, 7700 to 8200 for the central portion of the Manzanita Mountains, and 9600 to 9800 for the higher parts of the North Manzano Mountains.

The vegetation changes with rising elevation from short-grass plains in the western piedmont through pinyon-juniper woodland and western pine forest to northern coniferous spruce-fir forest in the higher part of the North Manzano Mountains. The Manzanita crest reaches only a few hundred feet into the yellow pine belt.

Surface water is scarce. Only Tajique Creek has living water for more than a mile. There are short flowing stretches in the north fork of Canyon Diablo (sec. 14, T. 6 N., R. 5 E.[1]), Hell Canyon, and Ojito Canyon. Several springs maintain flows for short distances.

General Geology

Structurally the area is relatively simple. A thousand feet or more of low-dipping Pennsylvanian sedimentary beds lie on a peneplaned Precambrian basement. The present mountains resulted from Tertiary block faulting in a zone several miles wide, and with a relative downthrow dominantly to the west. Such activity accounts for the Manzanita Mountains, whose even crest is a physiographic disconformity between a dissected east-sloping erosion surface and the steep western slopes resulting from dissection, retreat, and embayment of several fault scarps. During block faulting the Rio Grande Depression was filled with fluviatile clastics from the rising mountains; a few miles southwest of the mapped area these sediments are more than 4500 feet thick. Ground movements still occur within the Depression. The resulting seismicity is in general of low intensity, but, as recently emphasized by Northrop (1945), its frequency is exceeded in the United States only by those of California and Montana.

The North Manzano Mountains coincide with a previous, moderately high, east-facing ridge formed by high-angle thrusting from the west, possibly in Laramide time. The strata which formed this earlier eminence were the upper 2000 feet or more of Permian and Mesozoic beds and have since been eroded.

Precambrian metamorphic rocks beneath the flat Paleozoic unconformity record a third and older mountain group. Thick clastic and volcanic rocks were thrust north-northwest and intruded by granitic magma. The metamorphism is almost entirely of very low grade, and the rocks are here referred to the Proterozoic.

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[1] Locations throughout, where reference is made to the section grid, give Township North and Range East of the New Mexico principle base and meridian.

Darton and others have demonstrated that Ordovician to Pennsylvanian sediments are in conformable sequence southward across New Mexico. The entire column lies on peneplaned nonfossiliferous older rocks. Thus, the thrust-

sive bodies, both acidic and basic, also occur, and the largest have been mapped.

The Sevilleta formation was named by Stark and Dapples (1946, p. 1134) in the Los Pinos Mountains. The metarhyolites of the Man

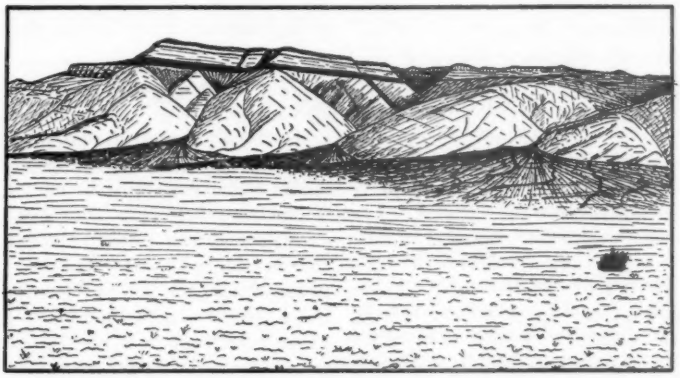


FIGURE 2.—NORTH MANZANO MOUNTAINS FROM THE SOUTHWEST
Bosque Peak on left. Sketched from a photograph.

ing and intrusion are pre-Ordovician and probably pre-Cambrian.

PRECAMBRIAN FORMATIONS

General Discussion

The older rocks are exposed in a narrow foothill belt along the west base of the range and in more limited areas in the southern and eastern part of the North Manzano Mountains. Since the general strike of bedded formations is approximately at right angles to the mountain front a natural geologic section is afforded. Six main units have been recognized and mapped. From apparent youngest to oldest these are: Ojito stock and related intrusives, Sevilleta metarhyolite, minimum thickness 4500 feet; Upper Metaclastic series, 4000 feet plus; Lower Metaclastic series, 7500 feet plus; Greenstone complex, including the Lacorocah meta-andesite tuff member, thickness unknown, certainly more than 2000 feet.

The Sevilleta and the Upper Metaclastics are conformable; an angular unconformity separates the two metaclastic series, and the relation of the Lower Metaclastics to the Greenstone complex is unknown. Numerous small intru-

zano Mountains are correlated with those of the Los Pinos district on the basis of lithologic similarity and position above heavy white quartzites. Since future mapping in the South Manzano Mountains may establish correlations between the metaclastic formations of the Los Pinos and those herein described, no new names are introduced at this time. The Upper Metaclastics of this report probably include representatives of the Sais quartzite, the Blue Springs formation, and the White Ridge quartzite of the Los Pinos Mountains, as defined by Stark and Dapples, and the Lower Metaclastic Series and Greenstone complex of this report probably represent new formations.

The thicknesses, relations, and distributions of each of the six units are summarized. The structures are reserved for a later section.

Greenstone Complex

The oldest rocks recognized constitute a thick series of predominantly greenish hue occurring in two districts characterized by poor outcrops, rounded slopes, and a rubbly mantle of relatively infertile soil. On either side of lower Hell Canyon, massive greenstone is an important constituent; north of Coyote Springs

chloritic schists predominate. Associated rock types include gray, black, and lavender slate, buff vaguely schistose siltstone, basic blocky metatuff, also slightly schistose, and gray andesitic metatuffs. The last named forms a distinctive member in the Hell Canyon district. No amygdules or flow structures have been found in the greenstones, although their volcanic origin seems probable.

A specimen from the NW¼ sec. 27, T. 8 N., R. 5 E., is massive except for pale, 0.5 to 1 cm. angular inclusions. It is fine-grained, decussate, with green hornblende 69 per cent, pennine 15 per cent, quartz and clouded feldspar 5 per cent, kaolinitic aggregates 10 per cent, and sparse hematite, pyrite, epidote, and exceedingly fine quartz mosaic. (All percentages in thin section descriptions are estimated.) The angular inclusions are chiefly unoriented muscovite, with a little oligoclase and quartz, and have clouded kaolinitic rims.

Many of the chloritic schists probably resulted from slight dynamothermal alteration of massive greenstone; others seem to have derived from fine-grained clastic sediments. The two types are not readily distinguished in the field. However, green schists suspected of sedimentary origin are as a rule silicified and very fine-grained.

As an example of the metaigneous variety a green, fine even-grained (average .05 mm.) chloritic schist from the SW¼ sec. 21, T. 8 N., R. 5 E., was sectioned. It is chiefly pennine with associated albite and pyrite. Calcite forms both discrete grains and narrow mosaic veinlets. Trains of minute quartz and very rare grainlets of clinozoisite (?) occur.

Another example from the Hell Canyon road near the east line of sec. 20, T. 8 N., R. 5 E., is mottled gray green and very fine-grained. It consists of sporadic 1 to 2 mm. insets in a matrix, 80 per cent, which has an average grain diameter of 0.1 mm. In the matrix are vaguely aligned albite subhedrons 45 per cent, quartz 15 per cent, minute muscovite 5 per cent, and a pale-green chlorite with birefrigence up to 0.012. Epidote grainlets, rarely a hematite flake, and small patches of cloudy kaolinitic debris are also present. The insets include epidote, with associated pennine, the two in many places forming confused aggregates whose outlines suggest former augite crystals. Formless calcite aggregates, quartz grains, and badly altered albite subhedrons also appear as insets.

Under the microscope representative specimens of the metasedimentary variety exhibit alternate wisps and bands of clear and much-clouded material. In some cases the bands are unresolvably fine-grained, in others they are mostly minute angular quartz grains with associated kaolinite, very fine-grained calcite, and, rarely, an albite grainlet. A few shreds of chlorite occur. In some specimens the quartz and albite include rather abundant rutile (?) needles. Quartz mosaic clots and wavy veinlets cross the schistosity.

Light-buff vaguely schistose bands within the Greenstone complex are a puzzling feature in the Hell Canyon district. The largest of these have been indicated on the map (Pl. 5). In secs. 21, 22, 31, and 32 of T. 8 N., R. 5 E., the trend of the bands parallels the local schistosity; in sec. 28, however, this is not the case. If the bands were bedding features their rapidly changing width and attitude would suggest an east-plunging syncline and a northwest-plunging anticline. The rock is probably a metasiltstone.

Under the microscope a representative specimen has a vague lamination depending on approximate orientation of sericite and of concentrations of cloudy decomposition products. Quartz grainlets predominate. Scattered unclouded albite grainlets and minute zoisite (?) subhedrons occur. Quartz, as in other sections examined, shows no strain effects. Some specimens have minor muscovite and ore grainlets. All are dominantly of silt grade.

An example from NE¼ sec. 29, T. 8 N., R. 5 E., is speckled and has a splintery fracture. It consists of 0.1–0.3 mm. angular quartz grains 25 per cent, slightly smaller feldspars, chiefly acid oligoclase, 20 per cent, and magnetite less than 1 per cent, set in a colorless unresolvably fine granular matrix containing cloudy kaolinitic patches, and very rare minute rutile and magnetite grainlets.

Slate interbeds in the Greenstone complex are lavender, gray, and in at least one instance black and carbonaceous. Lenses of the lavender variety a few tens of feet thick are common in the central part of sec. 23, T. 8 N., R. 5 E., and in the old Lo de Padilla Grant, 1½ miles northeast of Domingo Spring. Bedding parallels schistosity. Gray slates and phyllites crop out along the road in Lower Hell Canyon.

In the E. ½ sec. 29, T. 8 N., R. 5 E., there are slightly schistose basic to intermediate blocky tuffs. One of the few good exposures is in the cut made for the headframe of an old mine.

The Lacorocah metatuff member of the Greenstone complex[2] extends northeasterly across the NW¼ sec. 22, T. 8 N., R. 5 E., and into adjacent sections. It is light gray, crudely bedded, and weakly schistose, with clastic texture and rare or no quartz. Some beds are crowded with light-gray, slaty, flattened pebbles. A thickness in excess of 1000 feet is present between a fault on the northwest and a southeastern contact of undetermined character.

In thin section a representative specimen consists of equant 0.5–0.8 mm. sericitized acid oligoclase subhedrons, with minor quartz and magnetite. Veinlike, transverse bladed pennine separates many of the feldspars and seems to be attacking them. Clusters of minute clinozoisite and of fine quartz mosaic are rare. Sporadic small patches, deeply clouded and unresolvably fine-grained, may represent lithic fragments.

The member shows great diversity; it ranges from altered acidic lava, now quartz-albite rock, to a talc-tremolite aggregate presumably of very basic igneous origin. Such extreme types are not common.

Lower Metaclastic Series

Around the south end of the North Manzano Mountains the Precambrian consists of a great thickness of dominantly gray slates, phyllites, and schistose grits, in two series separated by an angular unconformity of 20° which extends diagonally across secs. 16, 20, and 30, T. 6 N., R. 5 E. The upper few tens of feet of the lower series is strongly silicified, and for 3 to 4 feet at the unconformity there has been essentially complete replacement by massive milky quartz. The beds above and below the unconformity are similar, except for a thick basal quartzite of the upper series. However, the older beds, cropping out to the west, are generally lighter gray and more arenaceous than those of the younger formation. Metamorphism in both series is of low grade, and the schistosity nearly or quite parallels the bedding. The fine-grained schistose grits in the upper part of the lower series are characterized by the abundance, in some laminae, of unoriented green, spindle-shaped, 4 to 10 mm. long concentrations of very fine-grained chlorite and magnetite. Their form suggests chloritoid metacrysts.

The older series is of indeterminate thickness, since its lower contact is the discordant Ojito stock. A narrow band of the series wraps about the east side of the stock and occupies a small discrete area in the outer foothills to the northwest. In both occurrences the material is heavily silicified, in striking contrast to its practically unaltered character along the south margin of the stock where the bedding dips away from, rather than toward, the intrusive contact.

The contact between the Lower series and the Greenstone complex is very poorly exposed. It has been impossible to verify the depositional relationship which has been used as a working hypothesis.

North of lower Hell Canyon a large area of metaclastic outcrop is here assigned to the Lower series. Since both contacts with adjacent formations are faults, and since the Upper and Lower series are lithologically similar, the implied correlation is somewhat gratuitous.

The metaclastics north and west of lower Hell Canyon are about 9500 feet thick. The beds are duller gray and coarser to the southeast, and probably at least the southeast third of the column is properly correlated. Unfortunately the evidence as to order of superposition is not entirely satisfying. Possibly both Upper and Lower series are represented. An unconformity of less than the 20° noted west of Comanche Canyon might have escaped notice.

Foliation and fracture cleavage are best developed in heavy bedded slates and phyllites to the northwest, where some hillsides reflect the sun as though treated with aluminum paint. In this district the attitudes of slaty cleavage and bedding often diverge slightly.

The metasandstones are mostly feldspathic, sorted, and of fine to very fine grade. Very rarely, as on the high spur ridge east of the center of sec. 16, T. 8 N., R. 5 E., there is a bed of very coarse to granuliferous sandstone, weakly schistose due to aligned muscovite flakes, and criss-crossed by quartz veinlets. Metatuffs have been recognized in sec. 18, T. 8 N., R. 5 E.

The weakness of the metamorphism is marked. A phyllite in the western part of the exposures, for

[2] Here named; the term is the Isleta (Tewa) equivalent of Hell Canyon.

example, is characterized in outcrop by roughly rectangular pits, 2 to 3 mm. across, from which calcite aggregates have weathered. The matrix is of quartz, 80 per cent, and sericite, with a few per cent each of hematite and magnetite, and a very little tourmaline. The calcite aggregates are clouded by kaolinitic matter and generally have narrow quartz-strain. An outcrop of this rock in the gulch bottom immediately southwest of the center of sec. 4, T. 8 N., R. 5 E., is distinctly gneissose. In thin section it differs from that just discussed in having a few per cent more quartz, 6 per cent less potash feldspar, 8 per cent of green hornblende, and an occasional titanite grain.

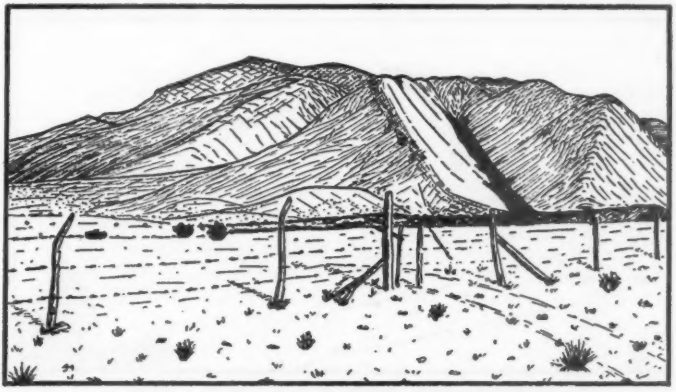


FIGURE 3.—SOUTH END OF NORTH MANZANO MOUNTAINS
Basal quartzite of the Upper Metaclastic series forms dip slope in center. Sketched from a photograph.

mosaic rims of somewhat coarser texture than the sedimentary quartz of the body of the rock. Few other specimens show calcite.

Higher grade metamorphism is suggested by small areas of a light-gray or dull pinkish-gray, massive to weakly gneissic medium-grained rock, the largest outcrop of which is in the SW¼ NW¼ sec. 8, T. 8 N., R. 5 E. A thin section has subhedral and anhedral microcline insets, 2 to 7 mm. in diameter, many of which grade outward into microperthite. These are set in a finer-grained matrix, with potash feldspars 20 per cent, sericitized acid oligoclase 45 per cent, quartz, in mosaic patches and individual grains, 15 per cent, green biotite 3 per cent, granular epidote aggregates 8 per cent, magnetite 2 per cent, pennine 5 per cent, and calcite 2 per cent. In the field this was considered an altered igneous rock. However, Dr. F. K. Morris concludes, from study of the thin sections, that it represents a clay which was first metamorphosed and then underwent extensive solution replacement. He notes that there is more calcite than could be yielded by other minerals of the rock, "even by more wholesale alteration than that here recorded," and that the habit of the calcite—in lens-shaped aggregates which parallel the vague schistosity in close association with the oldest minerals present—is not that of alteration calcite. Furthermore the biotite is strongly recrystallized, and the quartz is young, showing neither break nor

As an example of metatuff, a mottled red and green finely schistose rock exposed in the pediment in the NW¼ sec. 18, T. 8 N., R. 5 E., was sectioned. It contains 0.3 mm. grains of acid oligoclase (Ab_{88}) 40 per cent, thickly set in an obscure matrix dominantly of fine- and very fine-grained quartz mosaic, patchily clouded with kaolinitic matter. The matrix also includes epidote aggregates 20 per cent, pennine 4 per cent, iron ores 2 per cent, and a little muscovite. Another probable metatuff crops out about 180 feet west of the northeast corner of the same section. It resembles the specimen just described in the clastic appearance of its coarser fraction and in the thoroughly recrystallized quartz mosaic matrix. The accessories, however, are all somewhat coarser, and there is slightly rounded quartz as well as feldspar, chiefly albite, of sand grade.

Upper Metaclastic Series

From the angular unconformity at the south end of the North Manzano Mountains a thick metaclastic succession extends southeasterly beyond the limits of the area studied. The general dip is southeast. The lowest member is a white and gray quartzite about 400 feet thick, which has been separately mapped. Locally it has low-angle cross lamination which

confirms the evidence as to normal order of the beds given by the unconformity. This quartzite may well be a correlative of the Sais quartzite of the southern part of the range (Stark and Dapples, 1948, p. 1127).

Conformably overlying the quartzite is a comparable thickness of fine-grained chlorite schists and quartz-chlorite schists. These exhibit drag folding and shearing and are in places crowded with stout small lenses of vein quartz. To the southeast they grade into gray and greenish-gray chloritic slates and phyllites with occasional 10- to 15-foot beds of greenish very fine-grained silicinate sandstones. Lensing reefs of buff, massive, saccharoidal quartzite, which weather to narrow hogback ridges, are conspicuous in this part of the section. Higher, on the Comanche Canyon trail and in the west fork of Diablo Canyon, in sec. 15, T. 6 N., R. 5 E., red and gray "paper phyllites" appear, associated with red and brown, fine-grained schistose grits and siltstones. Outcrops in the north fork of Diablo Canyon are chiefly of gray slate.

In the Coyote drainage at the north end of the area mapped, metaclastic rocks aggregate nearly 2200 feet. Of this, a maximum of 1600 feet in sec. 8, T. 9 N., R. 5 E., consists of predominantly dark-red schistose grits, siltstones, and phyllites. Schistosity appears to parallel bedding, and there is superimposed plication and fracture cleavage in the fine-grained members. To the southwest the beds tongue jaggedly into gray, banded quartzites. The banding follows the original bedding; its darker grays correspond to concentrations of very fine-grained specular hematite. The boundary between the red and the gray parts of the formation, very conspicuous in the field, was mapped. The individual beds carry through;

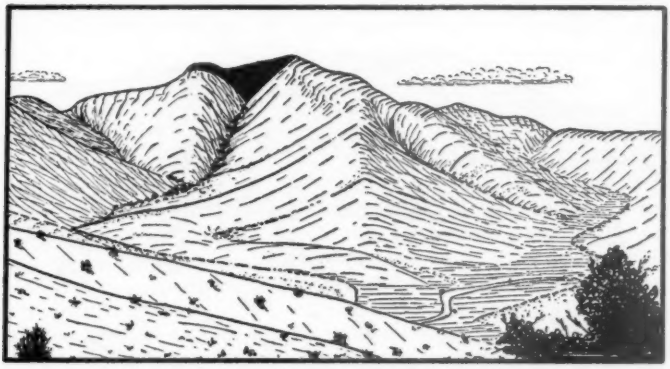


FIGURE 4.—VIEW NORTH ACROSS COYOTE CANYON

One and one-half miles east of the Springs. The skyline is notched by Precambrian red beds. Sketched from a photograph.

the darker-red sands and phyllites change within 10 to 20 feet into the darker-gray bands in the quartzite. Thorough silicification concurred with recrystallization of the pigmenting ferric oxide to produce the result noted. Locally, especially in the SW¼ sec. 12, T. 9 N., R. 4½ E, conspicuous segregation of minute hematite flakes obliterates the bedding. The rock is then gneissose, with strongly contrasting broad black and white bands which are wavy and lack continuity. Very locally, the darker bands are not due to hematite flakes; near a thrust in sec. 13, T. 9 N., R. 4½ E., the quartzite is blackened by abundant small needles of tourmaline.

Rare cross lamination in the sands and quartzites show that the series is right side up. At the northwest (toward the base) the exposures are limited by a fault, and the alteration has produced a continuous quartzite about 400 feet thick. This is particularly intensely silicified, and the bedding is obscured further by jointing and shattering. Another continuous white quartzite, 60 feet or more thick, overlies the main red-bed sequence, in the eastern part

of the exposures, thickening southwesterly and uniting with the expanding lower quartzite.

A silicified, gray, confusedly contorted and drag-folded quartz-mica schist lies conformably upon the quartzite. It averages 30 feet thick, increasing to 110 feet half a mile southeast of Coyote Springs. This forms the top of the Upper Metaclastic series, in this northern district.

The sequence which has just been described is thought to be stratigraphically higher than other metaclastics in the mapped area. Whether it constitutes a distinct formation or is correctly correlated as part of the Upper Metaclastics of the Comanche Canyon district is uncertain. The higher parts of that series contained some dark-red grits and slates. Similar materials occur sparingly in the northwestern parts of the metaclastics of Hell Canyon and as interbeds in the Sevilleta formation. The Coyote district rocks occur in normal sequence conformably beneath metarhyolite, similar, if not identical, with the Sevilleta formation of the Los Pinos Mountains. There also the metarhyolites are underlain by a heavy quartzite (the White Ridge formation).

Sevilleta Formation

The Manzanita foothills between Coyote Springs and, roughly, the north boundary of the Isleta Grant are developed in pinkish, vaguely schistose and gneissose metaigneous rocks, with minor associated dark slates, basic dikes and sills, and dikes of aplite, granite, and pegmatite. The Sevilleta rhyolite, named by Stark and Dapples (1946, p. 1134), is a 4500-foot formation in the Los Pinos Mountains which closely resembles and is almost certainly correlative with these pink schists and gneisses.

No bedding is evident in the Sevilleta formation in the Manzanita Mountains, but schistosity parallels the contacts of a few slaty interbeds in the central part of sec. 24, T. 9 N., R. 4½ E. If an uninterrupted homoclinal structure were to be assumed, the thickness of the Sevilleta would be of the order of 12,000 feet. Strike faults which might cause duplication of strata have been suspected, but could not be proven, both in sec. 30 and in sec. 31, T. 9 N., R. 5 E. An overturned syncline involving the entire outcrop belt, with its axis passing diagonally through sec. 31, is also a possibility. Stark and Dapples report a thickness of 4500 feet for the type locality of the Sevilleta in the Los Pinos Mountains, where, however, the upper contact is missing.

The metarhyolite, probably including considerable metatuff, is quite variable. Little-altered, massive, reddish aphanitic feldspar porphyry occurs north and east of the south quarter-corner of sec. 17, T. 9 N., R. 5 E.[3] Elsewhere, as in the N½ sec. 24, T. 9 N., R. 4½ E., flow structure is beautifully developed.

A representative specimen from the sec. 24 locality is nonporphyritic and consists chiefly of very fine granoblastic mosaic, over 80 per cent of which is quartz, with sporadic associated feldspar, mostly albite. Muscovite, 18 per cent, forms discontinuous trains outlining relict fluxion folds and lamination. There are two to three of these micaceous laminae to the millimeter, with intervening quartz granulite. Patches of kaolinitic debris and sparse hematite flakes occur. Lenses of quartz mosaic slightly coarser in texture (average 0.3 mm.) than the rest form many miniature saddle reefs.

A less-altered example has brownish chlorite and sericite streaks in a fine granoblastic quartz-feldspar aggregate. The estimated percentages of quartz, microperthite (including a little orthoclase), albite, chlorite, and sericite were respectively 25-45-23-4-3.

With advancing alteration narrow ill-defined coarser zones form 20 to 40 per cent of the rock and consist of sporadic 1- to 3-mm. insets of microcline-microperthite and microperthite with associated untwinned albite and smaller amounts of orthoclase and microcline. Minute magnetite grains are very rare. The larger grains are cracked and offset, and much of the quartz shows strain. Cracks along which there has been no offset are filled with fine quart mosaic; those in which the offset is pronounced, with felted muscovite aggregate. Many feldspar grains form small augen with quartz-mosaic pressure shadows (Pabst, 1931).

In general, the more altered metarhyolite is toward the south, where considerable thicknesses are medium- to coarse-grained gneiss, with a conspicuous lineation nearly paralleling the dip of the planar schistosity. The banding is very crude; strings of 1. to 2. cm. porphyroblasts of pink feldspar interrupt a dark, obscure, fine-grained aggregate.

[3] The quarter-corner is the midpoint of the section boundary specified; thus the west quarter corner is the midpoint of the west line of the section whose number follows.

In thin section a representative specimen shows feldspar porphyroblasts thickly set in a matrix which varies within the same slide from very fine quartz mosaic to fine aggregates of quartz with abundant apparently granulated plagioclase, muscovite, and biotite. Apatite, sericite, chlorite, and minute grains of epidote and magnetite also occur. Porphyroblasts include orthoclase, in part perthitic, and less commonly albite, which is kaolinitic, and, rarely, sericitic. Microcline occurs in 0.5 to 1 mm. clear grains, sometimes wrapped in biotite.

Rocks associated with the metarhyolite include metaclastics, dikes of aplite, coarse granite, and fine-grained altered basic rock. The intrusives are considered on subsequent pages.

Gray slate and fine-grained micaceous quartz schist form a discontinuous band in the Sevilleta from near the southwest corner of sec. 24, T. 9 N., R. 4½ E., to the SW¼ sec. 9, T. 9 N., R. 5 E. It is variably silicified and includes, in sections 24 and 19, a good deal of vaguely schistose rock, probably metatuff. The color ranges from pale green to dark reddish brown; megascopically massive to vaguely schistose, it is indistinctly granular.

Two thin sections exhibit cataclasis and a sericite aggregate, 75 per cent, which includes up to 10 per cent of chlorite in one specimen. Ragged 0.2 to 0.5 mm. strained quartz grains, 17 per cent, include rutile 4 per cent and magnetite 1.5 per cent, in both sections. In addition, the brown variety contains limonite-stained sericitized feldspar ghosts and about 2 per cent of lepidomelane; the green variety has 1–2 per cent of 0.2–0.3 mm. riebeckite prisms, many polysynthetically twinned.

In the W½ sec. 16, T. 9 N., R 5 E., on either side of upper Coyote Arroyo schistose grits and phyllites grade southeasterly from red to silvery gray. Small stout lenses of vein quartz are locally common. The assemblage closely resembles the Upper Metaclastic red beds 1½ miles to the northwest. It is cut off on the west by a granite intrusion near which it is darkened and silicified. Because no structure infolding or infaulting these metasediments into the Sevilleta could be found, they are mapped with that formation.

Intrusive Igneous Rocks

Silicic and intermediate.—Coarse-grained silicic and intermediate intrusive rocks crop out extensively on the western flank of the North Manzano Mountains and in the southern prong of the Sandia Mountains. Smaller bodies form dikes, most abundant in and adjacent to the Sevilleta formation, but present also in the Greenstone complex north of Coyote Canyon and along an ancient thrust fault midway of the west base of the Manzanita Mountains. Aplites are very common in the Sevilleta and in and about the granitic masses; pegmatites are relatively rare and inconspicuous. There is a single sill of rhyolite.

The largest granitic mass is here designated the Ojito stock.[4] The outcrop extends about 7 miles north and south (Pl. 5) and a maximum of about 2.75 miles east-west. To north, east, and south, the contact with the Lower Metaclastic series and the Greenstone complex is steeply discordant. On the west the intrusive rock passes under alluvial fans of its own substance and is faulted off within a short distance. On this side locally abundant xenoliths, rare in the interior of the stock, even at altitudes 2000 feet higher, suggest that the area of outcrop is not much less than the true limits of the mass. Apart from silicification, contact alteration of adjacent sediments is very slight, in contrast to the south margin of the much larger Sandia granite pluton.

The Ojito stock is light gray, massive, medium-grained and biotitic. No pronounced directional structures occur, even along its exterior contacts. Most if not all of the stock is metaluminous soda granite (Shand, 1943, p. 234 *et. seq.*) as illustrated by a representative specimen from the SW¼, sec. 28, T. 7 N., R. 5 E.

The thin section consists of quartz 27 per cent; sericitized oligoclase (Ab_{81}), many grains of which have clear orthoclose (?) selvedges, 44 per cent; weakly kaolinized orthoclase and minor microperthite 16 per cent; green biotite 4 per cent; hornblende, largely destroyed, and accessory sphene, apatite, magnetite, and zircon. Zircon inclusions in biotite have pleochroic haloes. Epidote, 6 per cent, replaces hornblende and feldspar; some grains look primary. Sericite, kaolinitic matter, rare zoisite, pennine after biotite, and calcite in untwinned

[4] El Ojito, literally the little spring, is the name of a permanent water source at the edge of the piedmont, and also of the drainage system to the southeast, which deeply and centrally exposes the stock. Elision and careless transliteration have resulted in the "Lolito" of some maps.

grainlets are secondary. The habit of the biotite suggests deuteric derivation from hornblende.

The rock appears uniform, but thin-section study indicates local gradation toward quartz monzonite. In such case, the potash feldspar is almost entirely microperthite, although the clear orthoclase (?) selvedges on oligoclase persist. One section shows striking sagenite webs of rutile in the biotite, which there amounts to over 10 per cent of the rock. Because of the microperthitic character of the potash feldspar, the rock designation remains unchanged.

Dikes and irregular masses of pink medium-grained granite occur in the Sevilleta formation, especially in its northern mile or so of outcrop and in the thrust zone which bounds it on the south. Only the larger of these were mapped. In places this granite has a distinct lineation and is not readily separable from the adjacent Sevilleta.

Pink granitic rocks crop out under and north of the quartzite phase of the Upper Metaclastic series in secs. 12 and 13, T. 9 N., and Rs. 4 and 4½ E. They resemble intrusives into the Sevilleta, except for cataclasis which increases toward the thrust contact with the quartzite, where marked schistosity parallel to the contact develops. The rock is locally almost mylonitized and is filled with secondary quartz laminae. At the northwest of the mapped exposures a small apophysis of granite extends about 60 feet into the Greenstone complex, narrowing and becoming finer-textured until, near its termination, it resembles a silicified rhyolite, with obscure flow texture.

Most of the northern granite is pinkish gray, fine- to medium-grained, and vaguely gneissose due to pinching and swelling laminae of pink feldspar, which alternate with finer-grained highly quartzose bands. There are sporadic larger lavender quartz grains (up to 2 mm.). Hand-lens examination discloses an apparent mortar structure.

Thin-section study of a representative specimen, taken 900 feet east of the NW. corner of sec. 13, T. 9 N., R. 4½ E., identified 0.5 to 3.0 mm. insets of strained quartz 10 per cent, and albite 8 per cent; the albite includes flecks of muscovite and rarely of calcite. The .06 to 0.1 mm. matrix has additional quartz 44 per cent, albite 11 per cent, orthoclase and microperthite 7 per cent, muscovite 14 per cent, calcite 4 per cent, wispy leucoxene aggregates 1 per cent, and sparse hematite and rutile, the latter included in quartz.

A finer-grained, massive, porphyritic phase predominates in the north-central part of sec. 13, T. 9 N., R. 4½ E. A thin section has about 21 per cent of 2.0 to 4.0 mm. phenocrysts in a 0.2 to 0.4 mm. ground. The phenocrysts are of quartz 6 per cent, in clear, equant, deeply embayed, cracked and slightly strained grains; and feldspar 15 per cent, including albite-oligoclase and minor orthoclase in slightly larger sericitized grains. The groundmass has quartz 28 per cent, and feldspar 51 per cent; the latter includes albite, orthoclase (?), microperthite, and microcline (rare). Obscure limonite-chlorite aggregates apparently follow crush planes. The groundmass is finely mosaic, suggesting recrystallization and possibly some silicification, and this rock may have been a porphyritic lava rather than an intrusive.

This northern body of granitic rock demonstrates a pre-thrusting episode of intrusion. The Ojito stock and most of the dikes of the central part of the mapped area are without marked strain effects.

Many small dikes in the northern part of the Sevilleta terrane are light-pink, conspicuously coarse, massive granite.

A specimen taken near the center of E½ sec. 24, T. 9 N., R. 4½ E., has strained quartz 30 per cent; fresh microperthite, including small grains of microcline, 42 per cent; sericitized oligoclase 23 per cent; muscovite and occasional biotite flakes, less than 1 per cent; and cloudy epidote aggregates. The quartz occurs chiefly in mosaic aggregates of grains less than 1 mm. across, but individual grains may attain 3 mm. width. The feldspars range from 6 mm. to 1. cm. in longer dimension. A very similar rock, also badly shattered, crops out in the SE¼ sec. 28, T. 9 N., R. 4 E., near the south end of a wedge between two Tertiary faults.

The outstanding example of an intermediate intrusive in the area is a discontinuous fin of quartz gabbro within the Ojito stock. This rock antedates its host, from which it is readily distinguished by somewhat finer grain and darker color.

A specimen from the ridge crest south of Ojito Canyon, in the NW¼ sec. 5, T. 6 N., R. 5 E., was studied in thin section. It is massive, fine-grained, hypidiomorphic; the light minerals average 1–2 mm.

in diameter, the dark minerals 0.5 to 1 mm. Included are quartz 11 per cent, and plagioclase 50 per cent, of which most is labradorite (An_{60}), but many of the larger grains have incomplete outer zones of andesine (An_{37}). Zoned hornblende 26 per cent, in places graphically intergrown with quartz, forms subhedrons and anhedrons, some few of which have kernels of diallagic augite. Shreds of biotite, 10 per cent, are pleochroic in pale yellow to dark brown, and locally crowded with magnetite grainlets. Apatite, zircon, and rutile are present in small quantity. Secondary products include kaolinitic matter and sericite in feldspar cores, sheaves of clinochlore associated with biotite, and a few fine aggregates of calcite associated with chlorite. The rock is noteworthy, both for the amount and size of the quartz in this association and for the long-continuance of mafic succession: augite → hornblende → biotite → clinochlore, which textural details suggest.

The margins of the quartz gabbro in many places are coarsely mottled and resemble rounded xenoliths locally common in the Ojito stock. In thin section the mottled rock is like the quartz gabbro except that the feldspar is chiefly basic oligoclase (Ab_{74}) and that biotite is lacking.

Rhyolite porphyry occurs only in a 12- to 16-foot sill which crops out conspicuously in secs. 17 and 18, T. 8 N., R. 5 E. Between 1- to 2-foot aphanitic selvedges, the rock is porphyritic. In view of the lack of metamorphism, the sill presumably postdated thrusting.

In thin section the 0.5 to 1.5 mm. insets, about 54 per cent, comprise quartz, orthoclase, albite (Ab_{98}), and muscovite, estimated at 23, 15, 15, and 1 per cent, respectively. The groundmass has an average grain size 0.1 mm. and includes orthoclase and albite 25 per cent, quartz 13 per cent, muscovite 5 per cent, and hematite 2 per cent. The rock is badly kaolinized. The occurrence of primary albite is of interest.

Pegmatite is relatively rare in the area studied. A few small, pink veins and stout lenses were noted in sec. 24, T. 9 N., R. 4½ E., but no work was done on them. Schorl is locally present.

Aplites of several types are common in and east of the Ojito stock and in the central portion of the Sevilleta formation where several large, white and pale pink, rather formless bodies were separately mapped. In general these are elongate parallel to the schistosity of the neighboring Sevilleta. Several examples were studied in thin section.

Aplites in the Sevilleta formation are mostly fine, even-grained, with marked orientation of muscovite and occasionally chlorite, of which only 1–2 per cent is present. Under the microscope, a specimen from the NE¼ sec. 36, T. 9 N., R. 4½ E., is a 0.25 to 0.5 mm. aggregate of sutured unstrained quartz 59 per cent, and a subhedral feldspar, chiefly albite (Ab_{98}) with very minor microperthite, 40 per cent. Many grains have clear selvedges (orthoclase?) about partly kaolinized kernels. Oriented wisps of fine muscovite aggregate, 1 per cent, a few clusters of minute epidote, and an occasional small riebeckite (?) grain indicate minor replacement metamorphism. The albite, however, is primary.

A deep pink aplite from the SW¼ sec. 18, T. 9 N., R. 5 E., is panallotriomorphic granular, with quartz 43 per cent, and slightly larger grains of feldspar 52 per cent. The quartz is seldom strained. The feldspar includes partly sericitized albite (Ab_{98}) 40 per cent, in 0.6–1.0 mm. grains, and microperthite which ranges in size up to 3.0 mm. There is 2 per cent of muscovite and associated lepidomelane which form short trains with a common orientation.

An aplite in the SW¼ sec. 35, T. 7 N., R. 5 E., has a groundmass, 90 per cent, which exhibits two preferred approximately perpendicular orientations of its mica. The groundmass minerals, averaging 0.1 mm. in diameter, include quartz 50 per cent; sericitized feldspar, probably chiefly orthoclase, 30 per cent; green biotite 10 per cent; and a little magnetite. The insets average 0.5 mm. in diameter and comprise quartz, biotite, and albite, which includes rounded quartz blebs.

Low-silica intrusives.—Four species of low-silica intrusives are present. Olivine gabbro forms several poorly exposed dikelike bodies within the quartz gabbro southeast of El Ojito. The rock is dark brownish green to black, massive, medium-grained.

A thin section of a specimen from the first bench southeast of El Ojito, in sec. 29, T. 7 N., R. 5 E., is poikilitic equigranular, with 1.0–1.5 mm. component grains. It consists of "labratownite" (An_{58-62}) 31 per cent, in twinned subhedrons; some of the smaller ones are included in diopside. Diopside, 31 per cent, forms anhedrons with very weak cleavage and patchy alteration to a uralite, *cf.* cummingtonite, 2 per cent, which in turn is giving place to brown biotite and magnetite on a small scale. Olivine, 28 per cent, forms deeply embayed 1.0 mm. anhedrons; it is highly magnesian with minor alteration to magnetite and antigorite.

Diabasic dikes and sills, mostly altered to greenstone, are sparse throughout the Precambrian terrane. Most are short and inconspicuous, but several are a quarter to three quarters of a mile in length and are mapped. Probably others are present but not distinguished in the very similar rocks of the Greenstone complex.

A number of small pluglike basic intrusions into the Sevilleta and the metaclastics of the Manzanita Mountains are medium- to coarse-grained, massive to vaguely schistose; many appear serpentinous in hand specimen. A few, which appear representative, have been studied in thin sections. The rocks probably originated by uralitization of pyroxenite.

Specimens of a conspicuous dike in the SE¼ sec. 21, T. 8 N., R. 5 E., and a smaller sill-like mass in the central part of sec. 24, T. 9 N., R. 4½ E., have green hornblende 90 to 95 per cent, which includes sparse, vague relic small pyroxene grains. The other minerals present are pennine, tremolite, calcite, and sericite, which appears to replace a few per cent of former feldspar.

A sample taken a few hundred feet northeast of the west quarter-corner of sec. 8, T. 8 N., R. 5 E., is massive and consists of frayed, relic, 1 to 2 mm. green hornblende, 20 per cent, in a patchy, finely felted tremolite and tremolite-chlorite matrix. The chlorite is a positive nearly colorless pennine. Magnetite, 6 per cent, in narrow trains and sporadic grainlets, and sparse apatite were noted.

A specimen from about 1000 feet west of the north quarter-corner of sec. 16, T. 8 N., R. 5 E., lacks hornblende. It is chiefly a massive, mottled green and colorless very fine aggregate, in which chlorite predominates over sporadic flakes of green biotite. Larger grains include calcite, mica, feldspar, magnetite, and apatite, 0.1–0.3 mm. in maximum dimension. The calcite is well distributed as individual grains and in veinlets and is often polysynthetically twinned. Mica, chiefly green biotite, is so strongly pleochroic as to be nearly opaque at the position of maximum absorption; many grains show sigmoid bending, others are interlaminated with muscovite (?). The feldspar is albite (Ab_{97}).

Dark, fine-grained andesite porphyry extends through the north quarter-corner of sec. 8, T. 8 N., R. 5 E., in and south of the zone of displacement of a Precambrian thrust. For a few tens to more than 100 feet from the main fault, the porphyry is a tectonite, with lineation approximately parallel to the south-southeasterly dip of the fault zone and to similar lineation in the neighboring Sevilleta formation.

In thin section the insets range from 0.5 to 2.0 mm. and are thickly set in an obscure 0.05–0.1 mm. groundmass. Twinned and variably saussuritized labradorite (Ab_{38-49}) with and without remnant zoning makes up about 31 per cent, and a green amphibole near riebeckite, commonly both singly and polysynthetically twinned, about 22 per cent. Much of the amphibole is crowded with magnetite grainlets. It occurs among the insets and also in the groundmass. Brown biotite, 39 per cent, in disoriented flakes some of which appear to wrap about the insets, is limited to the groundmass. Magnetite anhedrons, 3½ per cent, up to 0.8 mm. in diameter, are both included in and, more rarely, include amphibole. There is about 0.5 per cent of minute epidote and 2.5 per cent of quartz, the latter as mosaic and individual grains in the groundmass. Zircon and apatite are present as minor accessories, and a little kaolinitic matter and sericite as alteration products. The amphibole-magnetite relationship and the amount of biotite indicate some alteration.

Precambrian Structures

General Features

In this section are summarized all structures recognized as having developed during Precambrian times, except for the Ojito stock, the granite tectonics of which have not been studied.

An angular unconformity within the Precambrian indicates only mild disturbance in this district between the deposition of the two Metaclastic series. On the assumption of zero initial dips, the older beds in the SE¼ sec. 30, T. 6 N., R. 5 E., dipped 20° S. 7 W. when buried by the sediments of the upper series. This suggests that the stresses responsible had essentially the same orientation as those of the more impressive later diastrophism. The erosional surface of the unconformity was one of very low relief.

Schistosity, Fracture Cleavage, and Drag Folds

Development of schistosity varies with lithology, distance from large intrusions, and proximity to known or suspected Precambrian thrusts. The finer-grained clastics are most affected, tuffs are slightly less influenced, and

rhyolite flows still less, although here granitization was active. Basic intrusives are locally weakly schistose. Quartzose sandstones and red beds are schistose only near faults. In the Los Pinos Mountains Stark and Dapples (1946, p. 1130–1136) found schistosity directly related to the presumably Laramide thrusting. This is not so in the Manzanita and North Manzano Mountains where the single important Laramide (?) thrust is related only to a very weak and local schistosity in greenstones of the downthrown side. Nor does intense dynamic metamorphism extend more than a few tens of feet from known Precambrian thrusts.

The outstanding features of the local foliation are (1) relative feebleness of the alteration, and (2) the general parallelism of bedding and schistosity. Locally, as in the S½ sec. 17, T. 6 N., R. 5 E., both shales and sandstones seem unmetamorphosed. A few hundred yards northeast of the center of the adjacent sec. 20 a coarse quartzose sandstone closely resembles, for a short distance, sandstones in the lower part of the Magdalena formation. In general, however, sandstones have become schistose grits, and the shaly beds are slaty to phyllitic.

The slates and phyllites show (1) slaty cleavage; (2) minute plication of that cleavage; (3) drag folding; and (4) fracture cleavage usually related to the drag folds. Commonly the slaty cleavage occurs alone or is associated with only one or two of the others. In most places it parallels bedding, but in some areas, notably that north and west of the mouth of Hell Canyon, the strikes of the two diverge by 15° or more, and the dip of the schistosity is then slightly steeper than that of the bedding (Pl. 3, fig. 1). As a rule, bedding dips south-southwest, the foliation more nearly due south, where the two are at variance.

The relation of schistosity to bedding suggests that the Metaclastics series are uniformly in normal sequence. This is not confirmed by most other observations. Sandy rocks generally lack marked grade-size variation, even in thick beds. A few coarse sandstones and granule to small-pebble conglomerates near the center of the E½ sec. 16, and the center of sec. 17, T. 8 N., R. 5 E., offer unsatisfactory evidence; they have been silicified and badly interrupted by quartz veinlets, and their size variation is slight. However, in both places one is led to suspect that the section is overturned. Moreover, the entire series seems finer-grained to the northwest. No other criteria, such as those afforded by ripple marks, cross-bedding, or mud cracks, can be applied. The northwestern margin of the Lacoroch member of the Greenstone complex is quite surely right side up, as shown both by grade-size variation and channelling.

Drag folding on a small scale is common where slates and phyllites are interbedded with coarser-grained or more competent strata. With negligible exception the axes of the drag folds approximately parallel the direction of dip of the schistosity. The direction of asymmetry of the drag folds varies, suggesting open folding with a very steep pitch to the south-southeast and southeast. In the NE¼ sec. 8, T. 9 N., R. 5 E., such an open fold is recognizable.

Fracture cleavage is common, but by no means universal. It is essentially vertical and parallel to the dip of the slaty cleavage, rarely constituting an effective direction of weakness of the rock. Lineation in the metarhyolites and associated intrusives in general parallels the drag-fold axes of neighboring metaclastics, and presumably results from the same stress episode.

In some places a minute plication of the slaty cleavage occurs. Its axes generally trend nearly horizontally—*i.e.*, parallel to the strike of the schistosity—in which case, with a single exception, the downdip limbs of the plications are the "steeper". The exception is about 840 feet south of the center of sec. 16, T. 9 N., R. 5 E. Less commonly, a coarser plication parallel to the drag folding and fracture cleavage also occurs. The resultant cross-crumpling resembles that illustrated by Grubenmann and Niggli (1924, p. 458) and may best be seen a few hundred yards northwest of the south quarter-corner of sec. 8, T. 9 N., R. 5 E. Rarely, as in the S½ sec. 17, T. 8 N., R. 5 E., the only plication present is that which parallels the fracture cleavage.

Numerous joints in the Precambrian beds, some of them systematic, are with slight hesitancy referred to the Laramide or Tertiary disturbances and need not be considered here.

Folds and Flexures

Precambrian folds in the area are, with one very minor exception, possibilities, not demonstrated facts. A south-southeast dip of 50° to 80° characterizes both schistosity and bedding, nearly throughout the area mapped. Within the Greenstone complex and the Sevilleta formation evidence of bedding is scarce, and in Lower Metaclastic exposures north of Hell Canyon it is not certain whether the beds are in normal or reversed position.

Overturned folds may occur (1) within the Greenstone complex south of Hell Canyon, and (2) in the Sevilleta formation, as suggested by three circumstances—(1) Significant compression is proven by widespread slaty cleavage, and by thrust faults. (2) The exposed thickness of the metarhyolite is hard to credit unless duplication by thrusting or folding is present. (3) Both margins of the Greenstone complex appear to be in normal sequence, and in contact with younger rocks which dip southerly. While the northern contact is a steep thrust, the southern is possibly depositional. In view of the incompleteness of the evidence none of these is conclusive.

Flexing is common, though not of sufficient magnitude to outline distinct folds. It is especially marked in the metaclastics, south of the Ojito stock and in Commanche Canyon, where it impedes computation of thickness. In steep gulch walls, the same bed may change from a south to a north dip, and back, as one follows it down the dip. Similarly, the dip direction changes through the vertical along the strike in Commanche Canyon.

Thrust Faults

Several thrust faults mapped are Precambrian. The southernmost forms the contact between the Lacorocah member of the Greenstone complex and the adjacent metaclastics, passing diagnonally through sec. 21, T. 8 N., R. 5 E. Like the bedding on either side, the fault dips steeply to the southeast. It is recognized by a zone, 2 to 10 feet wide, of intensely sheared quartz schist with local lenses of hard, dark-red mylonite a few inches to a foot thick, which dip 75° SE. This material resembles, superficially, a slightly flow-textured lava.

A second thrust is inferred along the southern margin of the Sevilleta formation in secs. 5 and 7, T. 8 N., R. 5 E. A zone ranging up to a quarter of a mile wide exhibits advanced solution replacement and an intrusion of hornblende andesite porphyry. A gulch at the northern end of the zone exposes, for a short distance, a 6- to 14-inch seam of very fine-grained quartz which resembles alabaster in luster. The strike of the contact and seam parallels that of the schistosity on either side at N. 54° E., dip 75° SE. and on both sides also there is lineation at about S. 35° E. A minor fault in the Sevilleta is exposed by a prospect shaft a fourth of a mile to the northwest. Here the strike is N. 45° E., and the dip 85° to 90° SE. The inferred major thrust zone is interrupted by Tertiary high-angle faults in the western half of its trace. Probably this zone was the locus of displacements measurable in thousands of feet.

A shear zone, apparently corresponding to a small thrust, is exposed in the S½ sec. 24, T. 9 N., R. 4½ E. It is within the Sevilleta and is probably Precambrian. This zone dies out in the W½ sec. 19, T. 9 N., R. 5 E., and cannot be connected with a post-Pennsylvanian minor fault in the E½ of that section.

North and east of Coyote Springs, in sec. 13, T. 9 N., Rs. 4 and 4½ E., the contact between quartzite and granitic rocks below is an important south-dipping thrust. The quartzite is shattered, and the granite is flasered parallel to the contact for several tens of feet. The fault plane appears to have been folded by later and differently directed stresses and has the form of an anticline whose eastern flank is much the steeper. This feature is offset several times by later faults including, it is suspected, small flaws, but outcrops are poor. The steeply east-dipping contact between the quartzite and the Greenstone complex, which extends northeasterly as far as the east line of sec. 6, T. 9 N., R. 5 E., is obscured by silicification of both margins. Sharp changes of attitude across this contact suggest an extension of the thrust.

Mechanical Interpretation

Interpretation of the Precambrian structures seems to involve not less than three stress episodes: (1) Compression directed from the south-southeast and southeast caused slaty cleavage and thrust faults. (2) Cleavage was

plicated, with an asymmetry which seems to imply settling or relaxation of the rock mass toward the former direction of thrusting. Possibly this plication dates from a termination of lateral stress by widespread intrusion and cooling of granitic magma. For this explanation the Ojito stock to the south and the still larger mass of the Sandia granite to the north are conveniently located. (3) Later, east-west or southwest-northeast compression caused broad folds whose steep southeasterly pitch was inherited from the previous disturbances. Fracture cleavage, and occasionally a plication related to it, probably resulted from the same stresses. Since drag fold and fracture cleavage are best developed in the Manzanita Mountains, a district in which Laramide (?) strains are weakly represented, the compression responsible was probably older. No sufficiently severe disturbance is recorded in this part of New Mexico between Precambrian and Laramide time, and the features are, therefore, grouped with others of Proterozoic age.

This interpretation does not explain local divergences in strike of bedding and schistosity. Possibly the east-west compression was not later than, but in part concurrent with, the northerly directed thrusting. This might result from local wedging of the thrust mass between stable massifs, as of previously frozen intrusive matter to east and west. Exposures are lacking.

Paleozoic Formations

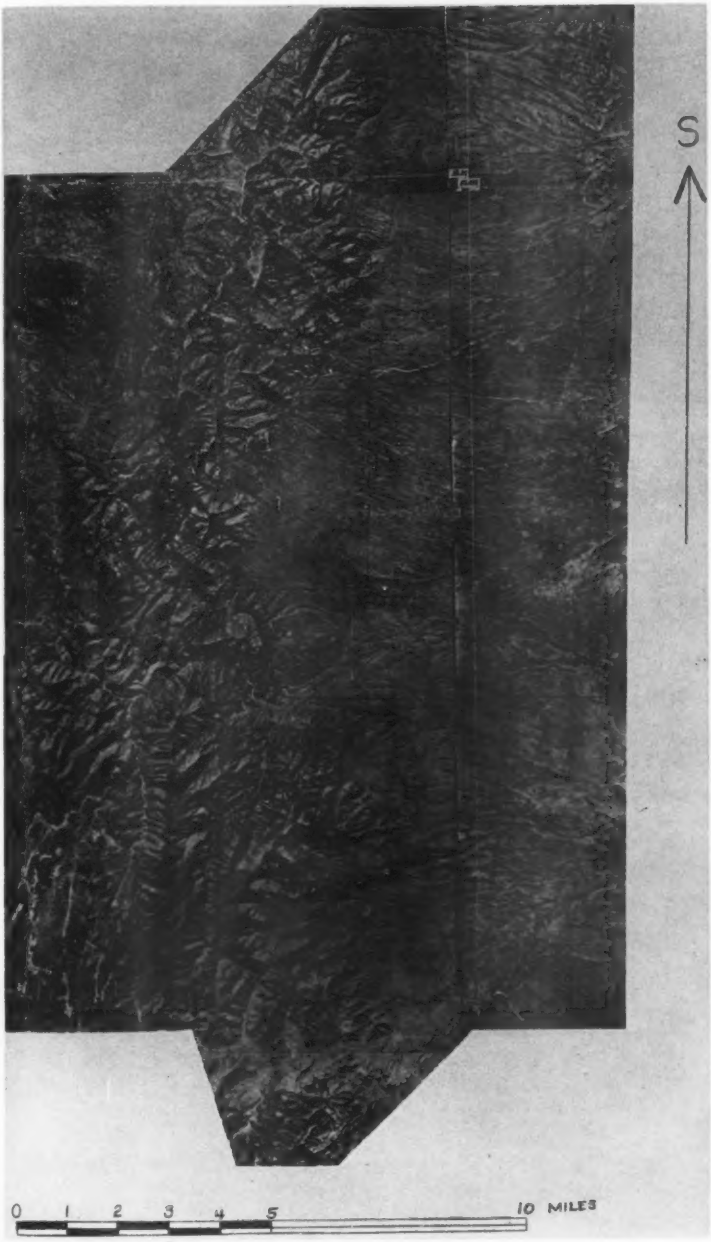
Pennsylvanian

The oldest Paleozoic beds in the Manzanita-North Manzano Mountains belong to the Magdalena group (Pennsylvanian). They rest depositionally on a peneplaned Precambrian surface of impressive flatness, except for rare elongate steep-flanked monadnocks. One of these, in secs. 6 and 7, T. 9 N., R. 5 E., is 250–300 feet high. Similarly, there is a little relief along the west side of the quartzite outcrop in the NW¼ sec. 2, T. 6 N., R. 5 E. (the east side has been faulted), and a similar condition is suspected from rapid inspection in the SW¼ sec. 16, T. 6 N., R. 5 E.

A pre-Magdalena formation was probably eroded before Magdalena deposition. Read and Andrews (1944) report a cherty limestone resting on the basement in the Sangre de Cristo Mountains, 50 miles to the northeast. This limestone, which also occurs in the northern part of the Sandia Mountains, Read provisionally refers to the Magdalena group (Read, Wilpolt, *et al.*, 1945). It is reported as about 15 feet thick immediately northwest of Ladrone Mountain, 55 miles to the southwest (Kelley and Wood, 1946). It is not known in the area here described, but in the W½ sec. 16, T. 9 N., R. 5 E., the lowest Magdalena bed contains, for a short distance, pebbles of chert certainly not of Precambrian derivation.

By extension of the initial significance of a name given by Herrick, the lower, dominantly clastic member of the Pennsylvanian succession, locally about 230 feet thick, has been known as the Sandia formation for many years. It consists of greenish-gray and brown, clayey siltstones and sandstones, with minor shaly beds and thin limestones. There are commonly one or two heavy beds of pebble conglomerate, one of them basal, but not universally present. The pebbles are predominantly subrounded white quartz, regardless of the local nature of the Precambrian. A single exception was noted near the E¼ corner sec. 8, T. 9 N., R. 5 E., where a thin basal sharpstone conglomerate consists of fragments of quartzite, locally derived. The Sandia is a slope-forming member.

The entire Magdalena above the Sandia, following a usage of Keyes, is known as the Madera limestone. It is predominantly cliff-forming limestones with intervening shales and minor sandstones. The Madera has been subdivided to emphasize the statigraphic significance of the initiation of growth of the Ancestral Rocky Mountains in the region to the east and northeast (Read and Andrews, 1944). Thus, the lower part of the Madera is called the "dark gray cherty member". This member is dominantly limestone and contains also gray shales and a few white quartzitic sandstones. In the Manzanita and North Manzano Mountains Wilpolt measured sections 400 to 450 feet thick (Read, Wilpolt, *et al.*, 1945). Read refers to the upper Madera either as the "arkosic limestone member" or the Canyoncito member". The former name is based on the markedly feldspathic character of the sandstone intercalations. The bulk of the member consists of limestones,



MANZANITA AND NORTH MANZANO MOUNTAINS
Photomosaic by Fairchild

FIGURE 1. MANZANITA FOOTHILLS
Sandia Mountains on skyline at right, and their southern prong (shadowed) in left distance. The Las Moyes—Las Seis confluence at lower left.

FIGURE 2. GUADALUPE AND MOSCA PEAKS
Note fault between the peaks, and stratum plain in foreground, right.

FIGURE 3. WESTERN FACE OF THE NORTH MANZANO SUMMIT
Densely brushed slope is underlain by the Sandia member of Magdalena formation.

VIEWS TO THE NORTH

many of them cherty, and gray shales. In some outcrops, silicate grains are disseminated in the gray limestone matrix, so the term "arkosic limestone" is accurate.

One such rock from the SE¼ sec. 11, T. 8 N., R. 5 E., contains about 15 per cent of subangular to angular noncarbonate grains, mostly of coarse sand grade. Quartz greatly predominates, but microcline, microperthite, and, rarely, a twinned plagioclase are also present. These grains are not in mutual contact, but scattered throughout the thin section, which consists primarily of rounded very coarse carbonate sand, mostly of fragmented Bryozoa. A calcitic cement results in low porosity.

Red siltstones, sandstones, and shales with Abo characteristics appear in the upper part of the arkosic limestone member. These become increasingly abundant, and the top of the member and the base of the Abo formation is taken on the uppermost limestone intercalation. Wilpolt's (Read, Wilpolt, *et al.*, 1945) measured sections show that the arkosic limestone member has a maximum thickness, in the area here studied, of 800–900 feet. In most places several hundred feet has been removed by erosion.

The part of the "Magdalena" which contains red beds is called Agua Torres formation by Stark and Dapples (1946, p. 1154), who present evidence for its Wolfcamp (Permian) age. As thus defined, the Agua Torres crops out only in the east flank of the north end of the North Manzano Mountains, in the area mapped. It has not been distinguished on the map (Pl. 5).

In view of the recency and thoroughness of the studies of Read and his staff, the writer has done no stratigraphic work. Read's subdivisions have been used in the mapping, but the contact between the two members of the Madera is subjective and might well be drawn 100 to 200 feet higher or lower in many places. This difficulty is well illustrated by the correlations shown on the published group of sections (Read, Wilpolt, *et al.*, 1945, section 5–12).

Excellent lithologic descriptions of the Magdalena, and a review of the zoning on microfauna advanced by Needham (1940), Thompson (1942), and Wilpolt (1942), are given by Stark and Dapples (1946). Most if not all of their discussion applies to the area here described. A summary of the regional aspects of Pennsylvanian sedimentation has also been published recently (Read and Wood, 1947).

Permian

Post-Magdalena formations are not present in the Manzanita-North Manzano Mountains, but several are exposed in the pediment belt to the west. Full sections of the Permian Abo, Yeso, and Glorieta formations are present in the Hubbell Spring district in secs. 9, 10, 16, and 21, T. 8 N., R. 4 E., where the lower part of the San Andres is also exposed. In the Los Ojuelos district on the Tome Grant the entire San Andres and Glorieta and several hundred feet of the upper Triassic Chinle formation are present. No angular unconformity is recognized between Magdalena and Chinle deposition.

The Abo formation, at the base of the Permian, consists of nearly 1400 feet of dark-red clays, siltstones, and sandstones, many of which are flaggy and ripple-marked. The upper surface of a dark-gray limestone is exposed in the SW¼ sec. 16, T. 8 N., R. 4 E., and the basal Abo resting thereon is a hard, ill-sorted, very coarse to granuliferous angular sandstone with abundant pink feldspar grains.

The top of the Abo and base of the Yeso formation was mapped, as is customary, on a sudden color change from reds which tend toward crimson (Abo) to definitely orange reds. The Yeso, which is about 1200 feet thick in the S½ sec. 9, T. 8 N., R. 4 E., consists of orange-red siltstones and sandstones with minor white sandstones and, in the upper part, especially in the Tome Grant exposures to the south, a few thin limestones.

The Glorieta and San Andres formations, which succeed the Yeso in that order, are best exposed in the Los Ojuelos district. The Glorieta is there about 125 feet thick and consists of yellow and buff, well-bedded friable sandstone; it commonly has low-angle cross-bedding. The Glorieta is sometimes considered a member of the San Andres formation (Read, Wilpolt, *et al.* 1945). Neither the large-scale cross-bedding nor the sparkling appearance, due to secondary growth of quartz grains, both of which are characteristic of the formation elsewhere, is well developed here.

The San Andres limestone, 90–100 feet thick in the Tome Grant outcrops, includes an upper 20–25 feet of limestone which is relatively massive and weathers yellowish brown; an unexposed 35-foot slope-forming member of reddish

clay or soft shale; and a lower 35 feet of thin-bedded, gray, locally cherty limestone, which in its lower part carries a poorly preserved gastropod fauna.

Mesozoic: Upper Triassic

The Chinle formation, incompletely exposed in the Los Ojuelos district, appears to be conformable on the Permian strata. Dark-red clays, siltstones, and sandstones, minor buff to brown sandstones, and a few thin limestones are present. South of the main Los Ojuelos spring, against the fault, a thin poorly cemented Chinle pebble conglomerate, in which desert-varnished jasper predominates, rests on red clays and is overlain by about 15 feet of thin-bedded limestone containing abundant small, fragile, unornamented ostracods and associated inconspicuous algal heads. It is supposed that the Chinle formation is here, as elsewhere, of freshwater origin.

Thick accumulations of Jurassic and upper Cretaceous sediments deposited over the entire district have been completely eroded from the Manzanita and North Manzano Mountains and their pediment zones.

Tertiary

Tertiary fluviatile sediments in the Rio Grande Depression, in places 4000 or more feet thick, have been described by Bryan and McCann (1937, p. 806–818), Smith (1938, p. 952–958), and Denny (1940a, p. 77–98; 1940b). Formerly collectively known as Santa Fe, a name given by Hayden in 1869, their lower portion has been split off as the Popotosa formation by Denny on the basis of an angular unconformity. In the area of this paper Tertiary beds are of limited distribution and are generally mantled by Quaternary debris. However, in secs. 27, 28, and 33, T. 9 N., R. 4 E., several low hills and a flat district to the east are formed of ill-sorted sandy coarse Tertiary gravels (fanglomerate). These are essentially impermeable, owing to an interstitial clay fraction which is about half calcite and half kaolinitic clay. Along their western margin the gravels are dragged to easterly dips as high as 48°, exposing a section 1100 feet thick; its extensions to east and west are masked by Quaternary gravels and colluvium which hide faults.

The lower part of these Tertiary beds has pebbles, cobbles, and, rarely, boulders derived from the Magdalena formation and lacks fragments of Precambrian derivation; the upper part has abundant Precambrian debris. The Magdalena crops out in a fault block 2 miles wide, less than a mile east of the gravels. Apparently drainage was established across this block while the gravels were accumulating.

To what extent the undisturbed gravels exposed in the lower Hell Canyon arroyo and in shallower draws in sec. 34, T. 9 N., R. 4 E., and to the south and southeast are Tertiary rather than Quaternary is not clear. The same doubt arises concerning the material penetrated by an Indian Service well near the center of sec. 11, T. 7 N., R. 4 E. It is reported that the well found only "granite wash" to its full depth of 239 feet. A water table was encountered at 109 feet. This "wash" may indicate either an old embayment of the mountain front or a graben behind the line of outcrops to the west.

Quaternary

The youngest materials of the district include pediment and alluvial gravels, blow-sand, colluvium and talus accumulations, travertine, valley alluvium, and occasional small landslides. Of these, only the blow-sands have been separately mapped.

Blow-sands are locally active, especially near watering places. Most blow-sands, however, are sufficiently old to have developed a soil profile. The most extensive are climbing dunes (Hack, 1941, p. 241) along the fault scarp of the southwestern part of Plate 5. Another stabilized dune area occurs east of Hubbell Spring. Active blow-sands are present south of lower Coyote arroyo.

At Coyote Springs, formerly known as Soda Springs, near the center of the N½ sec. 24, T. 9 N., R. 4 E., a terrace of travertine has been formed by waters emitting carbon dioxide, which rise under low pressure through the crush zone adjacent to a fault. Mass movement or cloudburst run-off repeatedly introduced angular quartzite fragments during terrace growth. Only one other spring in the district has deposited travertine,[5] although several, unlike

[5] 2000 feet N., 600 feet E., of SW cor. sec 11, T. 6 N., R. 5 E., too small to map.

Coyote Springs, issue directly from limestones.

Patches of black muck mark places of persistent ground saturation. Along the fault scarp in the Los Ojuelos district, where artesian springs have long been active, thicknesses of 8–10 feet are exposed at elevations up to 20 feet above present drainages, at several places. From one such place, Herbert Dick (personal communication, 1945) recovered two bison skulls. Artifacts and fire hearths of early Pueblo and possibly pre-Pueblo peoples are common in the vicinity.

Other types of Quaternary deposits are still forming. The shorter western drainage ways of the North Manzano Mountains debouch on steep fans many of which are 200 feet high. These characteristically support an open juniper woodland, while the peripheral pediment surfaces are merely grassed. Old pediments gravels extend 4–5 miles from the base of the mountains. Observed thicknesses are between 15 and 20 feet.

Post-Pennsylvanian Structures

General Statement

Since the flat contact between Pennsylvanian and Precambrian rocks is well exposed, later structures are usually easy to map. Very few can be more precisely dated by evidence available in this district. Normal and reverse faults are common, and there are a few related, small folds. In the wooded Magdalena terrane in the northeast part of the area small folds unrelated to known trends occur; time has not been available for mapping these.

Thrust Faults

The largest and perhaps the only really significant post-Proterozoic reverse fault in the area mapped is here designated the Padilla thrust, after the old Spanish Lo de Padilla Grant, within which its northern 3 miles lies. At the northwest this fault emerges from the alluvium of lower Canyon Las Seis; 6 miles to the south it passes into a steep monocline in the Magdalena formation; however, 1½ miles farther south along its trend a fault having the same characteristics arises and continues beyond the mapped area. Wilpolt has mapped a probably correlative fault segment 2 miles still farther south (Read, Wilpolt, *et al.*, 1945). East of Mosca Peak the vertical displacement on the Padilla thrust, which brings Precambrian on the west against Magdalena on the east, is at least 2000 feet. The Magdalena is crushed and dragged, in many places to the vertical, and locally, for short distances, it is overturned. Compression is also shown by an open anticline near the Fourth of July Spring. Detailed mapping proves that the fault, which is slightly sinuous, has a steep westerly dip. For 3 miles the fault trace parallels the steep margin of the Ojito stock, which probably controlled the location of the break. The tentatively correlated southern segment of the thrust is nearly vertical. Its throw, at the south line of sec. 15, T. 6 N., R. 5 E., is between 300 and 500 feet. Since Laramide thrusting has been established in near-by ranges it is reasonable to postulate the same general age for the Padilla thrust.

Minor thrusts within the granite in the rugged slope west of Guadalupe Peak are presumably also Laramide. Several of these features, roughly parallel, dip and are concave to the west. The most conspicuous underlies a flat-topped spur (Pl. 5). A klippe of granite, 50–100 feet thick, rests upon the fault plane, which at the west end dips 15° westerly. To the east the dip increases rapidly to about 55°. Along most of its trace the break is marked by a narrow oxidized band and minor cataclasis, but on the north side the hanging wall has been converted for about 18 inches into a cataclastic schist and dark-gray mylonite, with lenses of vein quartz. The schistosity dips westerly about 15° more steeply than the slickensided footwall. Great variation in intensity of crushing along the trace of this fault suggests a displacement of only a few feet, or at most a few tens of feet. The importance of the occurrence is that it confirms strong compressional stresses, since it has been argued by some that the west dip of the Padilla thrust resulted from slight rotation of an initially east-dipping gravity fault.

Similar, but smaller movements in the granite occurred on a number of roughly parallel surfaces. Several of these are conspicuous because of intensified weathering along crush zones. One, about three eighths of a mile north of that mapped, lacks schist or mylonite but

has cavernously weathered breccia 2 to 4 feet thick. Some of the openings extend back as much as 10 feet. Probably other similar zones of weakness are hidden by talus at the base of cliffs west of Guadalupe Peak.

The entire top of a knob seven-eighths of a mile east-northeast of Domingo Spring is a schistose cataclasite. Below this zone no well-defined fault plane can be traced for more than a few tens of feet. The best exposures are on the southeast, where a stout lens of milky quartz occurs near the base of the dynamically altered granite. The structure is complicated; lack of outcrop on the north renders its significance uncertain.

In the granite west of Guadalupe Peak and Mosca Peak there are several narrow high-angle shear zones, some of which probably represent minor thrusts. These strike N. 40°–65° E. and dip 55° to 70° most of them to the east. They weather in 6-inch to 1-foot relief, but can seldom be traced more than 50 or 100 feet. The zones, 1–4 feet wide, are flasered granite with seams of black slaty mylonite and, rarely, small lenses of white quartz vein matter. East-dipping examples occur in the dissected rock fan south and east of Domingo Spring, and an excellent 68-degree west-dipping example was noted about 1100 feet west-southwest of the northeast corner of sec. 32, T. 7 N., R. 5 E.

Gravity Faults and Probable Gravity Faults

Normal post-Pennsylvanian faults are numerous; only those with displacements exceeding 30 feet have been mapped. Throws of 100 to 300 feet are the rule, but several, not all exposed, have throws exceeding 1000 feet. All are relatively high angle; nearly all trend between north and northeast. Drag zones in adjacent sedimentary rocks are commonly restricted to the downthrown side, as would be expected in view of the limited thickness of flat-lying beds remaining on most upthrown blocks. There is no uniformity in direction of movement, but these faults are responsible for the present elevation of the mountains over the Rio Grande Depression, hence the net relative downthrow is to the west. Inasmuch as the top of marine Upper Cretaceous beds in the Rio Grande Valley is still near sea level, it appears that many of the normal faults which bound that Depression were the loci of actual, not merely relative, uplift.

The sinuous normal fault which branches from the sharp bend of the Padilla thrust has a throw of about 300 feet for much of its 4-mile course. The size of the northern of the two inliers of Precambrian along the fault in sec. 25, T. 8 N., R. 5 E., has been exaggerated for legibility on Plate 5.

One of the two faults shown on the State geologic map in the area of this report follows Hell Canyon arroyo for a quarter of a mile and is exposed in several prospect pits. This fault, dipping 70°–80° southeasterly, is relatively downthrown in that direction about 750 feet in the central part of sec. 23, T. 8 N., R. 5 E. Half a mile south of Hell Canyon its throw is less than 100 feet, but directly it splits into three branches, the eastern two of which outline a narrow graben. The easternmost branch dies out in a monoclinal flexure south of Las Moyes Canyon.

The fault which passes through the E½ sec. 20, T. 9 N., R. 5 E., branches repeatedly and involves a grabenlike sliver which contains the Sandia-Madera contact. It has a net displacement, relatively down to the west, of 150 to 250 feet.

About 380 yards south of the center of sec. 16, T. 9 N., R. 5 E., an exposed fault strikes N. 13° E. and dips 86° W. Striae suggest that the latest movement was a strike-slip. This is a reverse fault, since the east wall shows relative downthrow of 300 feet or more. However, the dip is very high, and the fault seems to be related to the normal faults of the district.

The "Coyote fault", which was responsible for the initial scarp of the northern two-thirds of the Manzanita Mountains, strikes N. 16° W., and dips 77° W. in a trench southwest of Coyote Springs in sec. 24, T. 9 N., R. 4 E. Here the throw is about 1300 feet. On the west the Sandia member of the Magdalena formation is dragged nearly to the vertical, against the quartzite of the footwall. A fault in the N½ sec. 1, T. 8 N., R. 4 E., may be part of the same structure. The Coyote fault, therefore, has an inferred total length of about 4½ miles and may extend still farther south. However, most of its displacement is probably transferred, via a short southeast-trending cross fault, to the break which, for a mile or so, closely follows the

line between Rs. 4 and 5 E. This, where seen, strikes N. 8 E. and dips 55° W.

The uplift of the range in this district is shared also by the eastern of three normal faults in sec. 7, T. 8 N., R. 5 E. South of the line of section C–C′ (Pl. 5) the graben between the western two of these three stands high, as a narrow ridge of Magdalena, with, near its terminus, a minor transverse monocline downstepped to the southwest.

For 1½ miles on either side of the mouth of Hell Canyon arroyo the sinuous marginal fault of the mountains is readily mapped and may be studied in prospect pits. A few hundred yards south of the arroyo it dips 74° W. and has a displacement of about 1500 feet. Very possibly this break is continuous with that bounding the North Manzano block on its west side. The latter is nowhere exposed; it is unmistakably implied by a battered fault scarp about 1000 feet high and, less certainly, by easterly dipping Triassic sediments 4½ to 5 miles west of the mountains. The precise location of the fault remains uncertain. In sec. 30, T. 6 N., R. 5 E., a minor branch fault trends off to the southeast where, 2 miles beyond the limits of Plate 5, upper Triassic red beds are in fault contact with Precambrian at the base of the South Manzano Mountains.

The logical line of division between the Manzanita and the Sandia Mountains is continuous to the north with the Tijeras Canyon fault shown by Read (Read, Wilpolt, *et al.*, 1945) and extends at least as far south as the south quarter-corner of sec. 28, T. 9 N., R. 4 E. Its total length, then, exceeds 12 miles. Throughout this distance it dips at a high angle; in the area shown on Plate 5 its relative downthrow is to the east, and it does not crop out. It brings the Sandia granite and its contact zone against Pennsylvanian limestones, Precambrian quartzite, and, at the south, upper Tertiary sands and gravels. Its displacement is unknown but presumably is measurable in thousands of feet.

Several faults branch southerly from the east side of the Sandia-Manzanita boundary fault. The easternmost is the Coyote fault. A second, branching off in the W½ sec. 13, T. 9 N., R. 4 E., parallels the master fault within a few hundred yards for more than a mile and shares its vertical displacement. A third fault, a quarter of a mile to the west, is downthrown to the west but dies out rapidly. The western downthrow, however, continues south on an inferred fault about half a mile still farther west. Thus a narrow horst extends for a mile south from Coyote arroyo. It consists of quartzite, metarhyolite, and the Sandia and lower part of the Madera members of the Magdalena formation.

The inferred fault passes approximately through the north quarter-corner of sec. 34, T. 9 N., R. 4 E. It is implied by the space relation of thick east-dipping Tertiary fanglomerates and low-dipping Madera limestones to west and east respectively. This relation appears to identify this fault with the Ojuelos fault.

The southwestern prong of the Sandia Mountains extends as far as the center of sec. 28, T. 9 N., R. 4 E.,[6] hence is not properly a part of the area covered in this paper. The "prong" consists of "marginal" granite, including roof pendants of epidiorite, sillimanitic nonschistose contact rocks and, at the south end, very coarse pink granite. The usual coarsely porphyritic texture of the Sandia granite is evident for a short distance along the north line of sec. 23, T. 9 N., R. 4 E.; elsewhere the texture is medium equigranular, and there is a southward-dipping initial platy structure and lineation which, in general, becomes more marked toward the south.

Quaternary displacements are evident near the western edge of the mapped area along the Ojuelos fault and, to the north, in the W½ sec. 22, T. 9 N., R. 4 E. At the northern locality a small fan built by Coyote arroyo is offset vertically about 25 feet. Although the scarplet is battered, the evidence is definite; it includes small fissure fillings of travertine due to a former artesian spring.

The Ojuelos fault developed during the Pliocene; Quaternary movements were relatively minor, recurrent, and apparently more recent at the south, in the Tome Grant, than in the Isleta Grant. Near Hubbell Spring, pediment gravels capping flat spurs are about 100 feet above the plain to the west, and Permian strata appear beneath them. The scarp is embayed and recessed; only the physiographic disconti-

[6] The margin of this mass is unfortunately labelled "Manzano Mts." on the new U.S.G.S. Albuquerque quadrangle. Lithologically, structurally, and geographically it is a continuation of the Sandia Mountains, and not part of the Manzanita, much less of the Manzano Mountains.

nuity and the presence of an artesian spring attest faulting. Continuing south, Hell Canyon arroyo has obliterated all evidence of faulting for several miles, but on the projected trace lies artesian Goat Spring. From sec. 32, T. 7 N., R. 4 E., southerly for 4½ miles, the fault trace is marked by stabilized climbing dunes built against aligned scarps which occur throughout this stretch, and beyond it to the south edge of the mapped area. In the southern 2 miles the scarps are about 110 feet high and formed in several stages. A 25-foot final uplift is most evident, since the flat floors of many canadas eroded in the eastern block after previous uplifts are as yet little trenched at that height. A bench level 50 to 60 feet above the plain to the west is sporadically present. Further evidence of faulting consists of artesian springs and permanent seeps near or on the trace of the fault. There is one each in secs. 5 and 8, T. 6 N., R. 4 E., and three main springs closely grouped at Los Ojuelos, on the Tome Grant.

The stage of dissection of the scarp suggests that most of the scarp-forming movement does not antedate the Recent. Far greater dislocations here or on buried faults to the west had occurred previously, and their surface expressions had been obliterated. Permian and Triassic strata are exposed east of the fault; to the west there are no outcrops, but the broad plain extending to the breaks of the inner Rio Grande Valley has been prospected for oil. The Grober #1 Fuqua well, 8.5 miles southwest of the mapped area, is stated by Dr. R. L. Bates (personal communication, 1946) to have drilled 4550 feet of Quaternary and Tertiary valley fill, 850 feet of upper Cretaceous, and 900 feet of probable Triassic, to a total depth of 6300 feet. There is physiographic suggestion of other Quaternary faults 2 miles southwest of Hubbell Spring, outside the area mapped.

Folds

Small folds of probable Laramide age include an asymmetric dome in the E½ sec. 1, T. 8 N., R. 5 E., and an open asymmetric anticline at Fourth of July Spring, in sec. 35, T. 7 N., R. 5 E. The latter was formed as a byproduct of the compression which caused the Padilla thrust immediately to the west.

Other folds of the district may be interpreted as drag along the upper Tertiary normal faults. In the W½ sec. 13, T. 9 N., R. 4 E., the Magdalena is thrown into an open south-plunging syncline by the opposed drags of faults on either side. A sharp bend on the western fault, in the SE¼ sec. 14, produced a narrow transverse anticline, which extends eastward into the synclinal area. Similarly, there is minor folding in sec. 20, T. 8 N., R. 5 E., possibly related to a bend of the adjacent fault.

PHYSIOGRAPHY

Comprehensive treatment of major physiographic features of the area is beyond the scope of this report. Indeed, no full explanation can be made until broad regional studies, such as those being carried on by the Harvard graduate group, under the direction of Prof. Kirk Bryan, are more advanced. Present knowledge and speculation are here summarized under four arbitrary headings: the North Manzano summit and east slope; the Manzanita Mountains; the North Manzano west flank; and the piedmont. Except for the Manzanita each of these is part of a physiographic division well represented in other segments of the range.

Much of the North Manzano summit upland is a stratum-plain remnant dipping gently eastward. It has been dissected by the headwater tributaries of Tajique Creek and Devils Creek, intermittent streams which descend eastward toward the closed Eastancia Basin. The stratum-plain remnant breaks off steeply on the east across the lower Madera limestones below which, at the north, an interrupted bench corresponds to the outcrop belt of the Sandia formation. In the saddle west of Fourth of July Spring this bench reaches the divide, and to the north the stratum plain has been essentially destroyed. The bench narrows and disappears to the south, where the Magdalena limestones flex downward suddenly in a monocline rendered obscure by poor outcrop and dense timber. Both Tajique and Devils creeks appear to be consequent, presumably upon the initial (Laramide ?) uplift of the range. Both have headward subsequent tributaries and are incised 100–200 feet at and beyond the east margin of the mapped area as a result of late Cenozoic uplift of the mountain block. The late mature

FIGURE 1. BEDDING AND SLATY CLEAVAGE
In central part of sec. 17, T. 8 N., R. 5 E. Outcrop faces west.

FIGURE 2. PEDIMENTS BEVELLING PERMIAN STRATA
In piedmont zone southeast of Los Ojuelos.

DETAILS

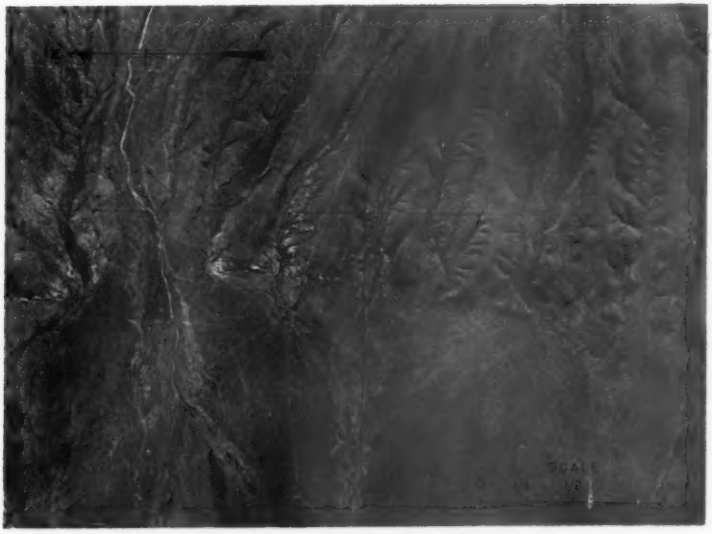


FIGURE 1. LOS OJUELOS FAULT SCARP
Photograph by Fairchild

FIGURE 2. WEST FACE AND SUMMIT UPLAND OF NORTH MANZANO MOUNTAINS
Bosque Peak in upper right; El Ojito canyon at left.
Photograph by Fairchild

AIR PHOTOS

surface from which the incision occurred is continuous to the north with the Manzanita upland.

The Padilla thrust determined subsequent stretches of valleys north and northeast of

FIGURE 5.—VIEW SOUTH-SOUTHWEST ACROSS THE MANZANITA UPLAND TOWARD THE NORTH MANZANO MOUNTAINS
Sketched from a photograph.

Mosca Peak. Silicified Precambrian beds adjacent to the Ojito stock form a broken ridge and have greater influence on relief. A high-angle northeast-trending fault near the southeast corner of sec. 10, T. 6 N., R. 5 E., lies west of the apparently subsequent valley of the north fork of Devils Creek which probably resulted from side-slipping of the stream down the east-dipping contact at the base of the Pennsylvanian east of the fault. The shattered zone along the fault, both where mapped and to the north along its projection, is the locus of strong springs.

The Manzanita Mountains form the high western edge of the Sacramento section of the Basin and Range province. Much of it is a late-mature erosion surface, possibly Pliocene. The structure is a gently east-sloping homocline diversified by minor faults and folds in upper Magdalena group sediments. The drainage pattern and specifically that of the steep western slope is thought to antedate the latest uplift of the range, since remnants of a former late-mature erosion surface occur along and near the present divide and between subsequent stretches of the rejuvenated tributaries of Hell and Coyote canyons. Many of the narrower spur ridges of the Manzanita west flank, however, are flat stratum-plain remnants cut at various horizons and bounded by precipitous limestone cliffs.

The Padilla thrust and the divide east of Mosca Peak mark the southern limit of the Manzanita segment of the range. Most of the run-off from the great cirquelike slope north and east of Mosca and Guadalupe peaks[7] has been pirated by headwater tributaries of Canyon Las Moyes; some, however, follows a subsequent canyon along the Padilla thrust into Canyon Las Seis. In the limestone country east of the thrust, and centering about 2½ miles northeast of Mosca Peak, the stream pattern strongly suggests a former centripetal drainage basin about 2 miles in diameter. At this late stage, proof would be difficult or impossible to establish, since the cavern system responsible would have been destroyed shortly after its presumably subterranean capture by Las Moyes Creek.

The west flank of the North Manzano Mountains is a composite landform developed in relatively massive granitic rocks. Approximately the lower 1000 feet is interpreted as an eroded fault scarp. However, the abruptness of the lowermost few hundred feet south of El Ojito suggests that that strip may be a fault-line scarp exposed by Pleistocene erosion. Above elevations 7000–7300 feet the slopes probably result from continuing deep erosion of a former mountain slope correlative with the initial (Plio-

[7] No evidence of glaciation has been found in these mountains.

cene ?) Manzanita upland, from which the North Manzano Mountains rose as a remnant Laramide ridge. There is no very convincing agreement in elevation of spur benches, indicative of recurrent uplift.

The uppermost 100–200 feet of the granite commonly stands as a nearly vertical cliff, maintained in places by sapping to crush zones along low-angle faults and elsewhere by current frost work. There are active talus tongues of huge blocks which afford welcome though strenuous approaches to the higher slopes in districts elsewhere densely brushy. At the south end of the mountain, in the E½ sec. 30, T. 6 N., R. 5 E., a quartzite dip slope (Fig. 3) has undergone dissection and waste removal probably diagnostic of a former more rigorous climate. The process responsible is no longer active, as is shown by the vegetated and healed gulch through which large volumes of quartzite debris were removed. It seems that only frost action could have wrought such rugged havoc and that a Pleistocene pluvial episode must be here recorded. A single rock fan has been recognized three-eighths of a mile southeast of Domingo Spring. Its surface has been almost destroyed by deep dissection, corresponding to recurrence of uplift after it had been formed.

The piedmont belt west of the mountains has three main components: alluvial fans, a broad pediment, and the terraced valley of lower Hell Canyon arroyo. Along the North Manzano front the fans, many of them 200 or more feet high, nearly or quite coalesce, and pediment growth must be due chiefly to lateral corrasion. Farther north, however, because of diminished relief and run-off, fans are small, if steep, and the knick line of interfan stretches results not from lateral corrasion but from the transition from rill to sheet wash at the foot of the slope, coupled with the tendency toward granular weathering of the Sevilleta formation. Where the Greenstone complex or the Metaclastic series forms the lower slopes, the knick line is less definitely developed. Since this northern stretch is a lower and less rugged scarp than the southern, the evidence of both supports the middle-ground explanation of pedimentation urged by Bryan (1936). The pediment zone appears to have an abnormally low slope to the west, but the possibility of tilting coincident with recent faulting renders this angle of dubious significance. Limestone cobbles in the surface layer of the gravels are strongly solution-facetted. The pediment surface is offset locally by Quaternary faulting. According to Wright (1946, p. 445) it correlates with Bryan's La Bajada (elsewhere Tio Bartolo) surface, one of several Pleistocene surfaces in the Rio Grande Valley. Its thalweg is stated to have been about 300 feet higher than the present bed of the Rio Grande.

The broad terraced valley of Hell Canyon arroyo is unique in the stretch of mountain front discussed in having had a sufficiently extensive water-shed to insure the larger floods reaching the Rio Grande throughout the entire period during which that master stream has been deepening its inner valley. This period dates from the destruction of the Ortiz surface, probably not later than middle Pleistocene, and quite possibly earlier (Bryan and McCann, 1938, p. 7–12). Quaternary alluvial terraces occur also along Coyote and Commanche arroyos. All are within 60 feet, and most are within 30 feet of present-stream elevations. Details of these and of such minor stream shifts as that of lower Coyote arroyo would burden the paper unnecessarily.

Climbing dunes along the fault scarp which offsets the pediment are stabilized and have a pedocal soil profile. Possibly they accumulated during the period of Recent aridity established by Hack in his work on the sand dunes of the Navajo country 220 miles to the west (Hack, 1941, p. 262–263).

Significant future physiographic work will necessarily include collaboration of archeologists and geologists. Remains of ancient "cultures" in this region include not only Pueblo and Basketmaker, extending back 2000 years, but also the vastly older "Sandia" artifacts, for which the type digging is only a few tens of miles to the north (Bryan, 1941).

Mineral Deposits

These data are based on brief inspection of the prospect pits, shafts, and adits in the district. Nearly all of these are in the Manzanita segment of the mountains. Fluorspar, lead, copper, and gold values have been sought; no mines are operating, and only one property appears promising.

In the area of Sevilleta outcrop, and to a lesser extent in the quartzite and Greenstone to the north, narrow discontinuous breccia zones along minor faults are locally either encrusted or cemented by coarse fluorite. Associated minerals are barite, galena, malachite, azurite, and pyromorphite. The faults along which these minerals have been deposited are mostly normal and of post-Magdalena, probably Tertiary, age. Repeated efforts over a number of decades have failed to develop a mine in any of these deposits, even under the stimulus of war prices. The deposits are of low grade; generally 15 to 30 per cent of a total breccia width of 2 to 5 feet is fluorite. The mineralized zones visibly pinch out within a very few hundreds of feet, both laterally and downward. In places partial solution and removal are evident.

Most effort has gone into a group of claims in sec. 31, T. 9 N., R. 5 E., but old prospects were opened to the southeast, east, and southwest. Some data on this district are recorded by Ross (1909). The prospect in the N½ sec. 7, T. 8 N., R. 5 E., is associated with a Precambrian thrust zone and shows more copper stain than most. A smaller group of similar deposits occurs in upper Coyote Canyon.

Many old prospects, especially in the Greenstone complex, were located on nothing more promising than a little iron or copper stain. There are very few quartz veins.

Probably the most promising property in the area studied is the Milagras group of patented claims in sec. 29, T. 8 N., R. 5 E. These include a narrow strip of placer ground, at the west, and one or more quartz lodes in the hills. Based on the remnants of an arrastra found over the ridge in Hell Canyon at the nearest surface-water source, there is a legend that the property was worked in Spanish days.

The Milagras claim was patented in 1876 and operated in the 1880's and early 1890's. The present owner, Mr. Alva H. Gunnell of Albuquerque, to whom the writer is indebted for this information, was informed by men associated with the undertaking that 1000 to 1500 tons of ore were run through a 10 stamp mill for a yield of $5 per ton in gold. The ore is a white to pale-yellow, highly porous quartz rock, from which, presumably, a high content of sulfides has been thoroughly leached.

On the Star claim, which adjoins the Milagras on the west, and is now held jointly with it, a shaft about 150 feet deep was opened in 1910. It is stated to have struck sulfides, and water, at that depth. Nine carloads of ore are said to have been shipped to El Paso and to have averaged $10 per ton in gold, and 9–28 per cent in copper. Like the Milagras, the property has been inactive for many years.

Immediately south of the Milagras claim a plexus of quartz veins is shown in very generalized fashion on Plate 5. Only the northern vein on which the Milagras is opened is of the porous quartz; the others are of dense chalcedonic vein material, varying from white to red which grades into silicified zones in the chloritic country rock. The appearance of the whole suggests replacement rather than fissure filling. Prospects about four-tenths of a mile to the south expose narrow bands in the country rock, marked by minor finely disseminated chalcopyrite and pyrite.

The local silication exhibited in the Milagras neighborhood occurs also in the Greenstone complex in the NE¼ sec. 28, T. 8 N., R. 5 E., north of Hell Canyon. There the silicified areas are small, oval, and consist of dark-red jasper. Only minor copper staining, such as is generally present in prospects throughout the Greenstone complex terrane, was noted.

Historical Résumé and Discussion

The geologic record of the Manzanita and North Manzano Mountains begins with the accumulation of several thousand feet of basic to intermediate tuffs and lavas, with intercalated clays and silts. These are Precambrian and from their relatively slight alteration are assigned to the Proterozoic. Many details of thickness, stratigraphy, and structure remain to be worked out. The accumulation is a monotonous series of dark greenstones and fine chloritic schists, with minor slates and weakly schistose tuffs, both lapillar and blocky. Pending further work the entire assemblage is grouped as the "Greenstone complex". All observed contacts are either intrusive or faulted.

Apparently younger than the Greenstone complex is a great thickness of metaclastic rocks, which, from their sorting, general fineness of grain, usual drab color, and well-bedded character, are thought to record marine deposi-

tion. No fossils have been found, and there is scant evidence of environmental type, beyond some cross lamination. Slates, phyllites, and schistose grits derived from clays, silts, and fine well-sorted sands, often slightly feldspathic, predominate in the lower part of this series, of which not less than 7500 feet is exposed between an eroded surface above and a discordant mass of granite below. Unlike the still thicker exposures northwest of lower Hell Canyon, this estimate probably does not include duplication of beds by faulting.

The lower clastic series and presumably also the material now forming the Greenstone complex were deformed, as shown by a low angular unconformity seen only on the southern part of the North Manzano Mountains. If the stresses were compressional they had a nearly north-south direction. Erosion bevelled the disturbed sediments, and renewed subsidence permitted the accumulation of more clastics similar to those below, but marked by more quartzitic sands and, at least in part, by typical red-bed development. The thickness of the later beds, the "Upper Metaclastic series" of this report, is in excess of 4000 feet. It is suggested that correlatives of the Sais, Blue Spring, and White Ridge formations of the Los Pinos Mountains are included. The succeeding formations appear conformable with the Upper Metaclastic series.

Volcanic activity on a grand scale now piled up more than 5000 feet of predominantly rhyolitic lavas and ash—the present Sevilleta formation. The vents have not been located. A dynamically altered pink granite to the north may represent an upper part of one of the magma chambers. During volcanic accumulation there were minor outbreaks of intermediate lava, and clastic sediments similar to those of the underlying series were deposited in limited basins. Probably as an aftermath of the vulcanism, and before extensive dislocation, masses of aplite with minor pegmatite and granite were intruded in and near the Sevilleta terrane. Very possibly it was at this time, also, that the Sevilleta was extensively granitized.

Several other pre-Cambrian rhyolitic sequences are known in this region. These include 1800 feet in the Franklin Mountains, 300 miles to the south (Richardson, 1909); and an unmeasured thickness in the Petaca district, 125 miles to the north (Just, 1937, p. 44). In the Zuni Plateau, 90 miles to the west, the writer has observed that much of the "granite" previously reported (Dutton, 1885) is actually rhyolite, some of it practically unaltered.

Still in Precambrian times, the rocks thus far noted were compressed from the south and southeast into steep folds and in places were thrust, mostly along steeply dipping planes. A slaty texture was imposed on the finer-grained sediments, and a weakly schistose to gneissoe one on sandstones and lava. Many of the basic lavas and tuffs of the Greenstone complex acquired weak schistosity, and the older granites north of Coyote arroyo were slightly flasered. Mylonite and a high degree of dynamic schistosity developed along some thrust zones, and along one there were irregular intrusions of granite and of porphyritic hornblende andesite.

Just (1937, p. 12–14) remarks on a general consistency of Precambrian structural elements having more or less east-west trends and northerly directed thrusting as marking a revolution which he terms "Puebloan". He notes further that the belt so characterized is very wide "possibly. . . extend(ing) from southern New Mexico to northern Colorado", and suggests that it may represent a western projection of the Killarney Mountain chain. Trends of similar thrust direction and general age occur in central Arizona, where they identify the Mazatzal revolution described and named by Wilson (1939, p. 1130–1134).

Following the compressional episode a stock of biotitic soda granite was emplaced in what is now the North Manzano Mountains. This intrusion, however, had been antedated by one of quartz gabbro which forms a broken irregular mass within the stock.

It is suggested that relaxation contingent on the intrusion and cooling of the stock and of granitic masses in neighboring districts was responsible for a plication of the slaty cleavage of the slates and phyllites. The axes of this superimposed structure are usually nearly horizontal, and the downdip flanks of the minute component wrinkles are almost always the "steeper". Probably the single rhyolite sill of the district was intruded during this same general time interval.

Possibly an episode of east-west compressional stress occurred later in the Cambrian

This hypothetical disturbance would account for open folding in the Metaclastic series. Since the beds retained a steep southerly dip from previous dislocation, the new folds pitch steeply to the south. They are accompanied by widespread development of drag folds and a fracture cleavage somewhere passing over into plication, nearly at right angles to that previously noted. This stress episode, if actual, ends the Precambrian record.

No lower Paleozoic events are recorded in these mountains. Instead, the older rocks were uplifted and eroded to depths of several thousand feet, attaining finally a peneplane, with occasional steep hogback monadnocks marking the locations of quartzites in the Metaclastic series. The first Paleozoic sediments to be deposited on this surface were marine cherty limestones, subsequently removed, except for chert fragments in the basal Pennsylvanian sediments, which are also marine. Presumably this history is one of repeated epeirogenic emergence and submergence. Remnants of the older limestone, still undated, are reported in neighboring districts.

Throughout Pennsylvanian, Permian, and much of Mesozoic time the Manzanita-North Manzano district was part of a large region undergoing recurrent sinking and sedimentation. Ground movements were epeirogenic. Probably the lower Triassic, the lower Jurrasic and certainly the lower Cretaceous were times of nondeposition, corresponding to long stands of the region above, but not far above, sea level.

Pennsylvanian sediments (Magdalena formation) are dominantly marine; the seas, judging by recurrent thick limestones, were shallow, warm, and bordered by lands initially of low relief. A few sandstone and many shale beds show that there were near-by lands. The basal few hundred feet of beds (Sandia member) are chiefly clastic, laid down during the gradual submergence. The succeeding cherty limestones and shales (Madera member) record the initiation of uplift of the Ancestral Rocky Mountains, as noted by Read *et al.* (1945), by the appearance in their upper portion of feldspathic sandstone, as contrasted with the quartzitic sandstone of the lower portion.

In earliest Permian time, continuing sedimentation accompanied by occasional slight regional downwarping built the sediments above sea level, at first tentatively and interruptedly (Agua Torres formation) and then uninterruptedly (Abo formation). This condition is shown by red-bed accumulations, which contain a sparse flora and vertebrate remains. The latter, however, have not been noted in the district reported. The Abo formation has characters suggesting flood-plain deposition from large low-gradient rivers flowing from neighboring uplands which were experiencing warm-humid to subtropical climates. Later Permian formations record a succession of changes both here and in the contributing uplands. The Yeso formation is both sandier and less well sorted than the Abo, indicating more spasmodic, possibly seasonal, deposition. In neighboring districts where the formation is better exposed, thick beds of gypsum and the presence of salt-crystal impressions in soft gray shales indicate that one reason for this change in regime was the shift to an arid climate. During Yeso time also, again in near-by districts, thin marine limestones appear. The Glorieta sandstone, above the Yeso, is extensively cross-bedded; in part it is probably of dune origin. At the top of the Permian, the San Andres limestone records another marine incursion, but the character of the intercalated clastics suggests no marked climatic change.

Only a fraction of the upper Triassic Chinle formation redbeds is exposed in the area mapped. Flood-plain deposition under an arid climate is indicated by a number of features. Thin lacustrine limestones, rich in ostracods, may constitute a new record for the Chinle. From previous studies it is known that thick Jurassic continental and upper Cretaceous marine and continental beds blanketed the area although they are no longer present.

Strong compression directed from the west, probably of Laramide age, is the next recorded event.[8] A high-angle thrust produced a steep

[8] The study of this area adds nothing to the precise dating of this compressional episode. According to other workers (Read and Andrews, 1944) the Sangre de Cristo Mountains to the northeast were mildly deformed in post-Mesa Verde—pre-Galisteo time (late Eocene-early Oligocene), and the Nacimiento Mountains, 50 miles to the north-northwest, were strongly thrust after "the deposition of the Torrejon formation of Paleocene age. Movement . . . may have continued intermittently during Oligocene times" (Northrop and Wood, 1946). *See also* the excellent discussion by Stearns (1943, p. 314–316).

east-facing ridge more than 2000 feet high on the site of the present North Manzano Mountains. Concurrently a similar but larger ridge rose at the site of the South Manzano Mountains. In these mountains the strata on the upthrown side have easterly dips which only locally exceed 5°. This condition contrasts with the Sandia segment of the range, which was interpreted by Darton as the result of Tertiary block faulting of a great anticline, the eastern flank of which constitutes the present eminence (Darton, 1922, p. 218–219). Thus the southern and central parts of the range (Los Pinos, South Manzano, North Manzano) yielded to Laramide stress chiefly by thrusting; the northern segment, as it appears, by folding. Lack of recognized strain in the intervening Manzanita segment may have resulted from its protected position behind the southern end of growing Sandia anticline.

A reasonable, if decidedly hypothetical, sequence of upper Tertiary and Quaternary events in this district can be pieced together from the observations recorded herein and from conclusions reached by Bryan, McCann, Denny, and Wright in their published studies of the Rio Grande Valley.

Regional elevation began in the middle Tertiary and was accompanied by block faulting which outlines the Rio Grande Depression. The faults responsible were dominantly north-south. More than 4000 feet of Miocene-Pliocene continental sediments accumulated in the Depression, which was bounded on the east by growing fault scarps near, or within a few miles to the west of, the present Ojuelos fault and its inferred north-northeasterly extension beyond Hubbell Spring. The scarps were battered and eventually buried by rising alluvium, now the Santa Fe formation, which is over 1000 feet thick against the northern part of the line, as shown by deformed gravels in secs. 28 and 33, T. 9 N., R. 4 E. Canyons cut in the eastern block were backfilled and obliterated. Late in this period of deposition, Precambrian rocks were exposed to erosion on the east side of, and probably as a result of uplift on, the Coyote fault. This is shown by changing character in the deformed gravels. Since the elevations of the still older Manzanita upland and its dissected counterpart east of the North Manzano Mountains are comparable, upper Pliocene uplift on the Coyote fault was probably matched to the south by uplift along it and other faults near the base of the mountains.

Aggradation was halted near the close of the Pliocene at a level 200 feet or so above the present piedmont surface. Successive incisions of the Rio Grande then resulted in the development of two erosion surfaces, probably in the lower Pleistocene. The earlier Ortiz surface is not known to be present in the area of this study. The second, La Bajada, was cut as a pediment several miles wide across Tertiary and pre-Tertiary strata to a position near the present base of the range. During this same time the fault which bounds the Sandia block on the southeast was active; Precambrian rock was brought up on the west against the Santa Fe gravels. The abruptness of the lower slopes of the southern part of the North Manzano Mountains suggests their Pleistocene origin, either by rejuvenation along a fault at the base of the range or as a fault-line scarp due to lowering of the piedmont zone.

Finally, probably in early Recent time, minor displacements occurred within and on the margins of the Rio Grande Depression. Of these the only local evidence is that of the low scarps of the Ojuelos and Hubbell Spring districts, and at the mouth of Coyote arroyo, sec. 22, T. 9 N., R. 4 E. Evidence of shifting climates during early Recent time has already received brief comment.

As one stands on the high summit of the North Manzano Mountains which is, at least hypsometrically, a culmination of the long history just discussed, one's attention is apt to wander from the immediacy of that record. From that vantage point the sharp horizon is 70 to 90 miles distant. It includes generous segments of the Colorado Plateau, with the Tertiary volcanic pile of Mt. Taylor, the Nacimiento and Sangre de Cristo prongs of the Southern Rocky Mountain, and, to the east, the Pecos section of the Great Plains, seen across the narrow and ambiguous Sacramento section of the Basin and Range Province. The mountain underfoot is characteristically a "basin range", yet its similarities to adjacent districts are not less striking than its differences. One sees that "the characteristic features of the Plateau Province extend far east of the Manzano Mts." (Lindgren, 1910, p. 163). In

the Manzanita upland Pennsylvanian strata have an elevation greater than they attain anywhere in the Plateau. If the Southern Rocky Mountains are characterized by Laramide thrusting, so too are the North and South Manzano Mountains and the Los Pinos Mountains. Nor can those three be distinguished on the basis of upper Tertiary and Quaternary block faulting. The same character may be claimed for the west face of the Sangre de Cristo Mountains and for the Nacimiento Mountains. Further, the Tertiary and Quaternary vulcanism of the San Juan and its New Mexico appendage, the Valle Mountains, is continued, though in very diminished form, in the basaltic cinder cones and small shield volcanoes within the Rio Grande Depression, which is a part of the Basin and Range salient. The Sacramento section of the Province reaches the present crest of this basin range in the Manzanita Mountains, and closely resembles both the Plateau to the west and the Great Plains Pecos section to the east. To find a strictly distinctive feature one is forced back to the Pennsylvanian and Permian, when mountains long since worn low and buried rose in the Southern Rocky Mountains, the Colorado Plateau, and the Great Plains, but not, in sofar as known, in this northeastern basin range.

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2801 Muir Way, Sacramento, Calif.

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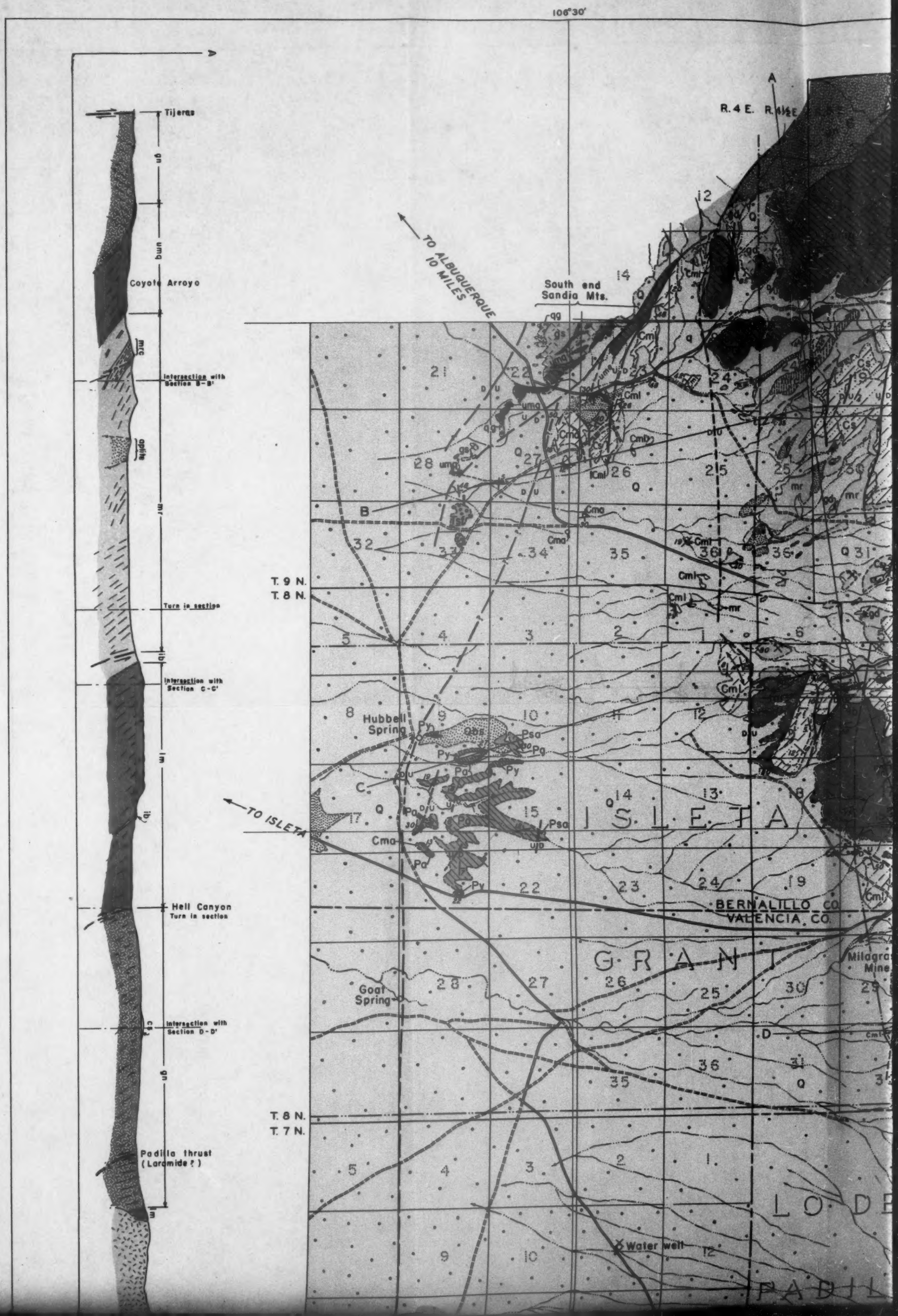

106°30'
A
R. 4 E.
R. 4½ E.
TO ALBUQUERQUE
10 MILES
South end
Sandia Mts.
Tijeras
Coyote Arroyo
Intersection with
Section B-B'
Turn in section
Intersection with
Section C-C'
Hell Canyon
Turn in section
Intersection with
Section D-D'
Padilla thrust
(Laramide?)
TO ISLETA
Hubbell
Spring
Goat
Spring
Water well
BERNALILLO CO.
VALENCIA CO.
ISLETA
GRANT
LODE
PADIL
Milagra
Mine
T. 9 N.
T. 8 N.
T. 8 N.
T. 7 N.
B
C
D

106°25'

106°20'

35°00'

34°55'

T. 10 N.
T. 9 N.

T. 9 N.
T. 8 N.

BERNALILLO CO.
TORRANCE CO.
T. 7 N.

BERNALILLO CO.
VALENCIA CO.

R. 4 E. R. 4½ E. R. 5 E. R. 5 E. R. 6 E.

A

B'

C'

D'

TO ESCABOSA

Milagras Mine

Moyer

Cañon Spring

Tijeras Arroyo

Arroyo Bonito

Arroyo Torreon

Arroyo Tijeras

Arroyo Tijeras

Arroyo Tijeras

LODE PADILLA GRANT

TOME GRANT

BERNALILLO CO. / TORRANCE CO.

VALENCIA CO. / TORRANCE CO.

T. 8 N.
T. 7 N.
T. 7 N.
T. 6 N.

R. 4 E. R. 5 E. R. 6 E.

Padilla thrust (Laramide?)

Turn in section

Klippe (Laramide?)

Intersection with Section E-E'

Cordova Ranch

Water well

Domingo Spring

Guadelupe Peak

Mosca Peak

El Ojito

Ojito Canyon

Fourth of July Spring

Bosque Peak

Tajique Ranger Sta.

Los Ojuelos Spring

TO TOME

Comanche

A'

E

E'

Elevation 5000

Elevation 3000

unconformity

0 1 2 3 4 5 MILES

SCALE

Coyote fault

Intersection with Section A-A' and turn in section

Sol Somete Canyon

Canyon Bonito

Manzanita Grade

B

B'

C

C'

Q Qu Qg Tsf Tsa Tsl Tsc Tsb Tsv Tsr Tsp Tsg Tsm Tsd Tsh Tsw Tsq Tsi Tsn Tsk Tsx Tsy Tsz Tsa Tsb Tsc Tsd Tsf Tsg Tsh Tsi Tsj Tsk Tsl Tsm Tsn Tso Tsp Tsq Tsr Tss Tst Tsu Tsv Tsw Tsx Tsy Tsz

Cml Cma Cs Cu Cv Cw Cx Cy Cz

Pg Psa Py Pv Pw Px Py Pz

Rc

p€ mr um im gn gr ib ig ia ic id ie if ih ii ij ik il im in io ip iq ir is it iu iv iw ix iy iz

upper Metaclastic series

lower Metaclastic series

PROTEROZOIC

mr: block schists (interbedded with graphitic metaquartzite)

um: slates, phyllites and pebbly grits

im: quartzites; feldspathic quartzites; schists of flandin schist zone (basalt crosses flandin schist zone)

lm: schistose grits, phyllites and slates

gn: Greenstone complex (greenstone, chlorite schist, chloritoschist and basic metalavas; lenses of metaschists and chloritoschist sometimes dipping)

gr: Lacroosh intrusive: biotite metatuff member

strike and dip of bedding

strike and dip of foliation

strike of vertical foliation

normal fault

thrust fault, sawteeth on upthrown side

abandoned mines

spring

LODE

PADILLA

GRANT

TOME

GRANT

Water well

Domingo Spring

Guadelupe Peak

Mosca Peak

El Ojito

Ojito Canyon

Fourth of July Spring

Bosque Peak

Tajique Ranger Sta.

Comanche

Ranch

VALENCIA CO.
TORRANCE CO.

BERNALILLO CO.
TORRANCE CO.

T. 8 N.
T. 7 N.
T. 7 N.
T. 6 N.

R. 4 E. R. 5 E.
R. 5 E. R. 6 E.

34°50'
34°45'

0 2 3 4 5 MILES
SCALE

PROTEROZOIC

upper Metaclastic series

lower Metaclastic series

mrc: black schists (interbedded with Sandia metarhyolite)

um: slates, phyllites and schistose grits

umq: quartzite; includes sillimanitic schists of Sandia contact zone (small crosses indicate brecciated zone)

lm: schistose grits, phyllites and slates

gn: Granestone complex (granestone, chlorite schist, intermediate and basic metatuffs. Buff metarhyolite and bleached granestone stippled.)

gl: Lacroosh intermediate metatuff member

strike and dip of bedding

strike and dip of foliation

strike of vertical foliation

normal fault

reverse fault, T on upthrown side

abandoned mines prospects

spring

B B'

Coyote fault

Intersection with Section A-A' and turn in section

Sol Somete Canyon

Elevation 5000

C C'

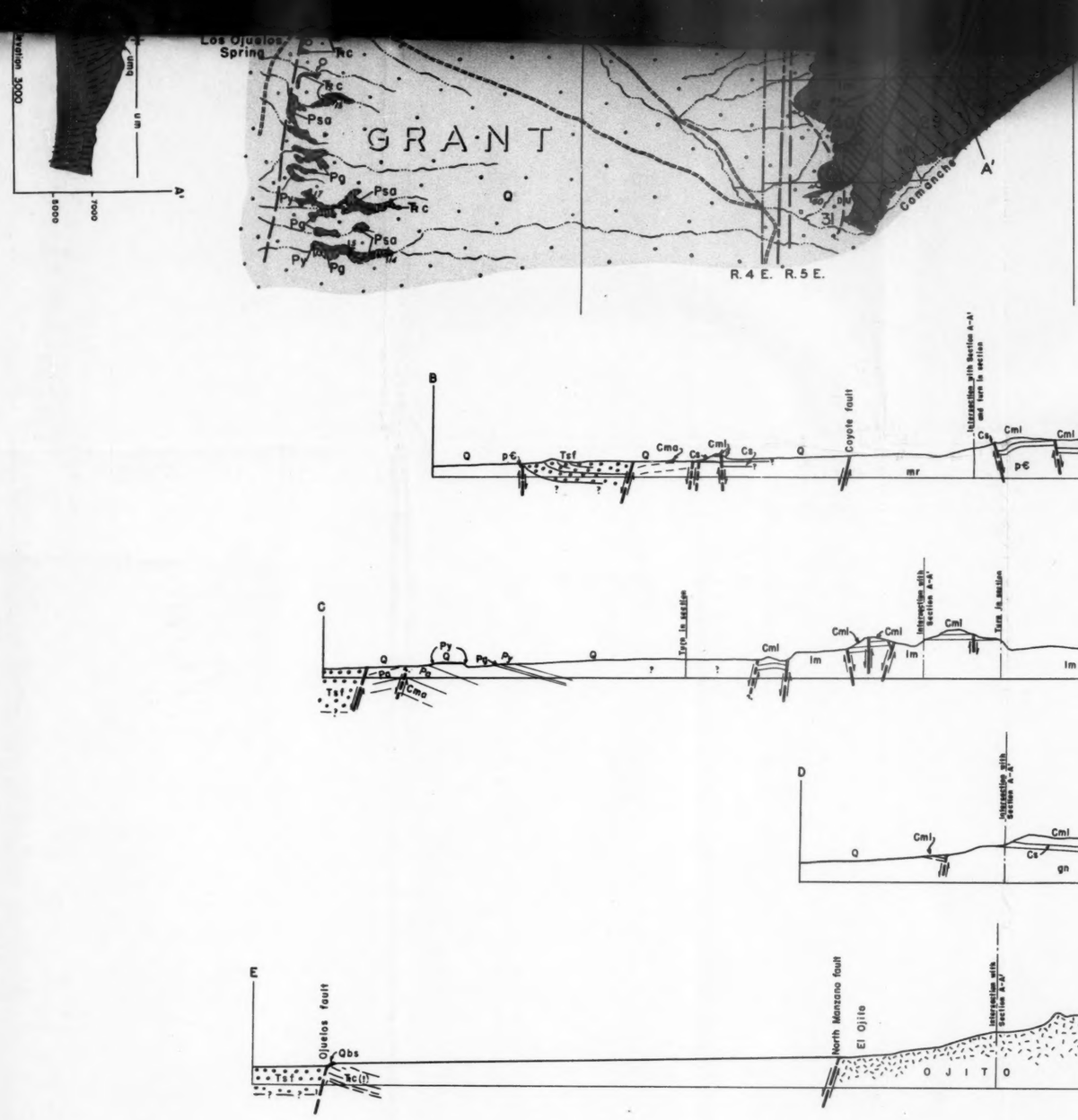
Intersection with Section A-A'

Turn in section

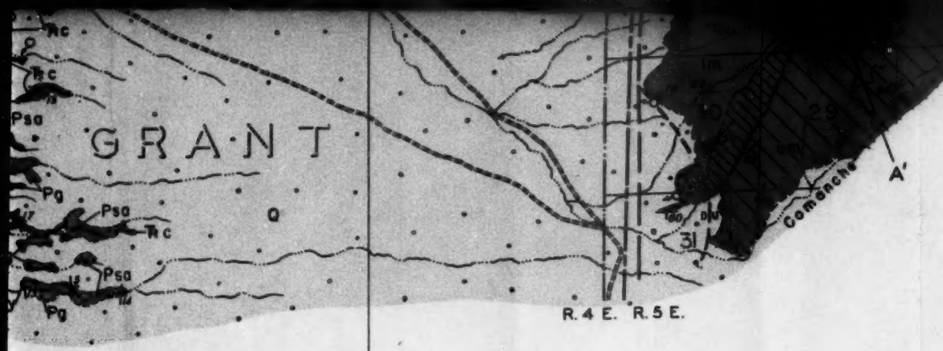
Canyon Bonito

Manzanita divide

Elevation 5000



GEOLOGIC MAP AND STRUCTURE SECTIONS, MANZANITA AND NORTH MANZANO MOUNTAINS, NEW MEXICO



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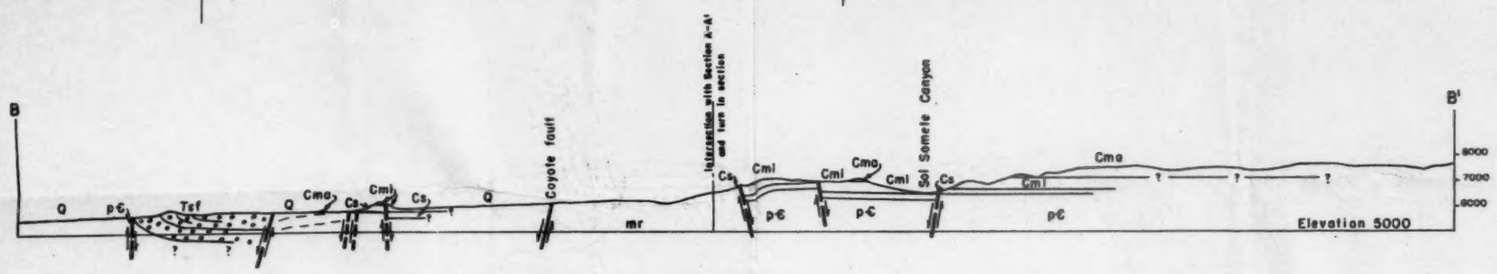
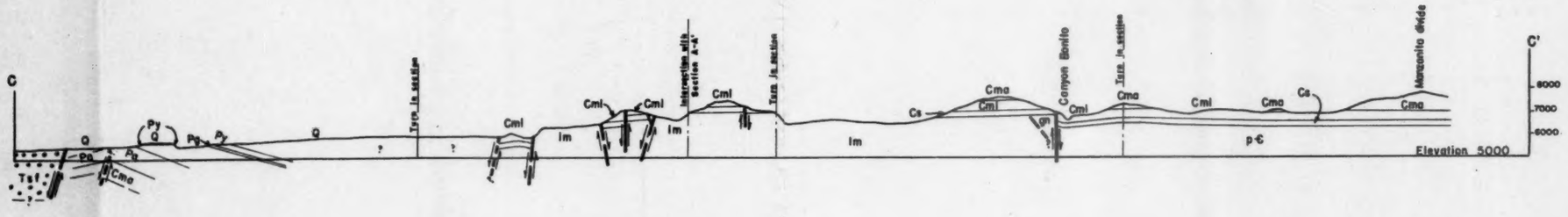
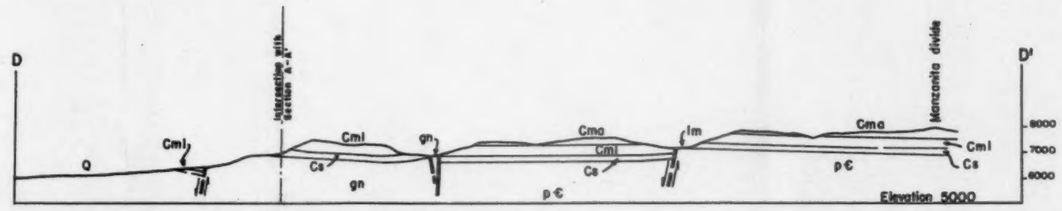
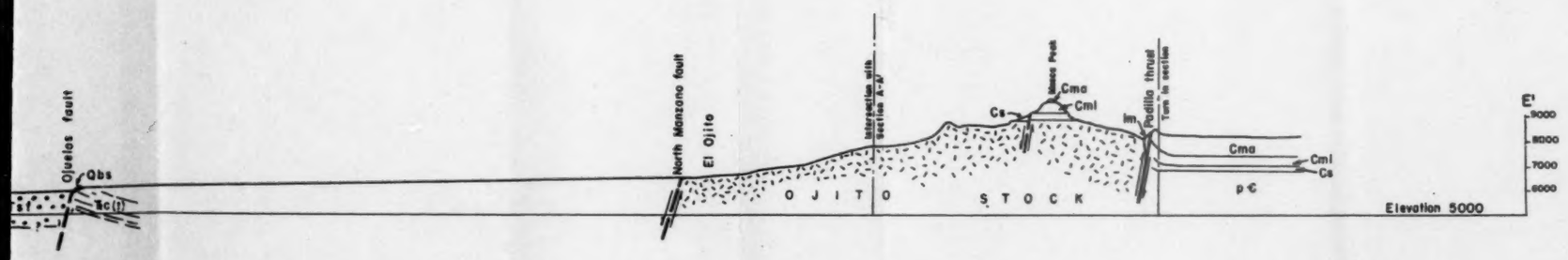
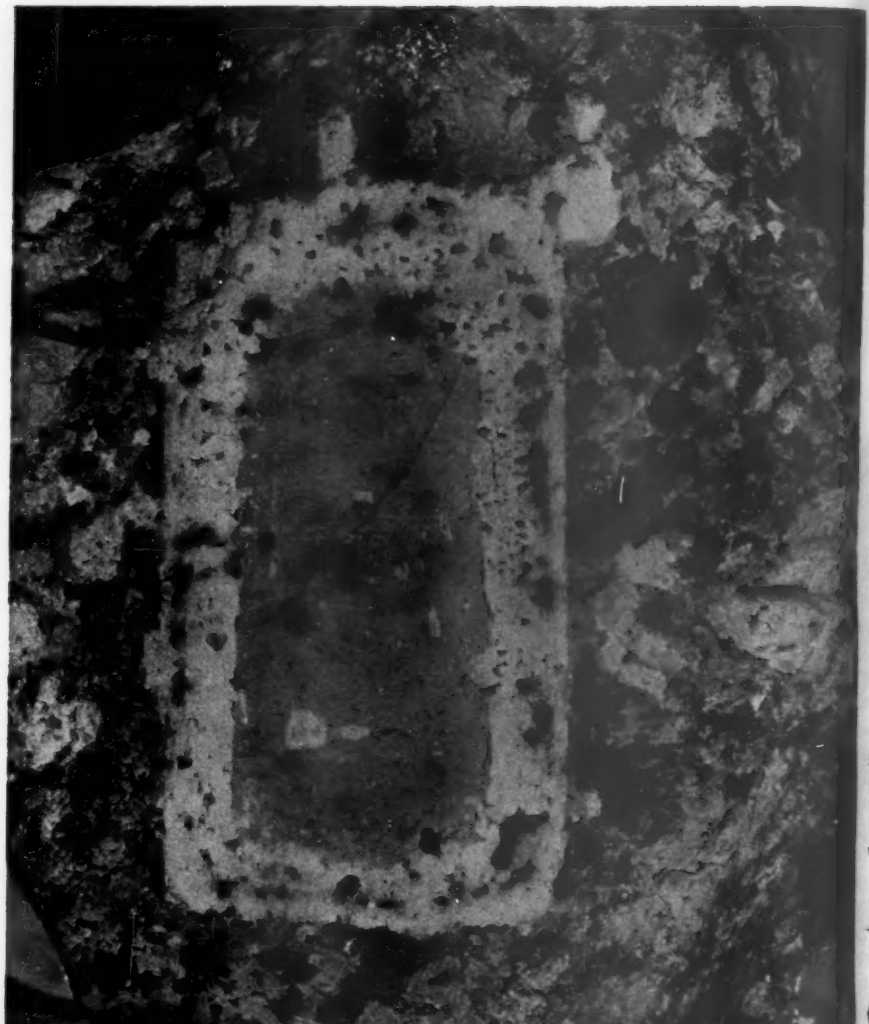








LOGIC MAP AND STRUCTURE SECTIONS, MANZANITA AND NORTH MANZANO MOUNTAINS, NEW MEXICO



INSET OF MICROCLINE (GRAY) WITH MANTLE OF PLAGIOCLASE (WHITE)
Found in granite of Table Mountain, Cape Town, South Africa. Magnified 2 diameters.

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Short Notes

HISTORY OF A FELDSPAR CRYSTAL

A Contribution to the Granite Problem

By S. J. Shand

The Table Mountain granite of Cape Town, South Africa, is a coarse-grained, porphyritic biotite granite with large insets of potash feldspar; we may call this feldspar microcline, but it is actually, on a minute scale, microcline-microperthite. The contact of the Table Mountain granite with the siltstones of the Malmesbury series has been known to geologists since Charles Darwin drew attention to it more than a century ago. For an accurate petrographic description of the granite and its contact phenomena one should refer to the recent work of Walker and Mathias (1947). A remarkable feature of the contact is a zone of streaky migmatite in which there are large microcline insets identical with those in the granite. Walker and Mathias formed the opinion that the granite, as a whole, is "not the product of granitization in place", but that the microcline insets in the granite as well as in the migmatite were generated by potash-rich solutions circulating through the solid rock. The main argument in favor of the second conclusion seems to be that, since the large microclines in the migmatite obviously grew in a solid matrix, the similar crystals in the granite must have grown at the same time and under similar conditions. Upon reading this conclusion, the writer was reminded of a remarkable specimen in his possession, which was collected near the contact on Kloof Nek, some 25 years ago. This specimen shows a single, exceptionally large inset composed of a core of microcline surrounded by a broad mantle of plagioclase. The full significance of the specimen was not realized at the time, belief in "granite magma" being then almost universal among petrologists, but the writer kept the specimen and used it in his lectures to illustrate zonal structure in feldspar. He was now led to make a closer study of this remarkable inset, and for this purpose the surface was ground flat and stained by the beautiful method of Gabriel and Cox (1929), which consists in exposing the rock to the fumes of hydrofluoric acid and then dipping it in a strong solution of sodium cobaltinitrite. The effect is to stain the potash feldspar a fine golden yellow, while plagioclase remains dead white and other minerals are unaffected. By this means the relation of potash feldspar to plagioclase is made delightfully clear, both on the specimen and in the photograph (Pl. 1).

The exposed surface of the inset is the basal cleavage plane, and the two nearly straight edges are the traces of the second pinacoid. The "ends" of the microcline core have a ragged appearance, but this does not seem to be a consequence of corrosion. If the little embayments were due to corrosion of microcline by an albitic liquid then they should be filled with plagioclase, but many of the embayments contain quartz grains. It seems logical to conclude that these irregular traces result from rapid growth of the crystal in the direction of the *a* axis at a time when little crystals of quartz and plagioclase were already present in the matrix. The straightness of the (010) and ($0\bar{1}0$) traces is explained by slower growth in the direction of the *b* axis. The entire crystal measures 58 by 31 millimeters (2.3 by 1.2 inches), and the microcline core 47 by 19 millimeters (1.9 by .76 inches) (maxima). The ratio of areas is there-

for very nearly 10 plagioclase to 8 microcline, and, whatever thickness one may assume the crystal to have, the ratio of volumes is about 2 to 1.

In the microcline core there are a few idiomorphic inclusions of plagioclase. These are so small, in relation to the giant microcline, that one can hardly doubt that microcline began to grow first. There are also some problematic inclusions of quartz intergrown with biotite; these may be fragments of country rock, but one cannot be sure of this, and the point is unimportant. The microcline crystal grew to a large size and then suddenly stopped growing. On its faces, here and there, films of quartz were deposited. These films are from .02 to .2 mm. thick, and some of them are in continuity with grains of quartz in the matrix of the rock. The thicker portions of film are recognizable in the photograph, where they look like biotite scales but are a little lighter in color.

The plagioclase mantle now began to form, and it continued to grow until it formed the greater part of the inset. During its growth it enclosed many separate grains of quartz and scales of biotite. These seem on the whole to increase in size from within outward, and their average size is distinctly less than that of similar grains in the matrix of the rock, so we may conclude that quartz and biotite were growing in the matrix all the time that plagioclase was being deposited. Yet it is remarkable that, although the matrix contains more microcline than anything else, not a single grain of microcline is enclosed in the mantle.

The plagioclase mantle having grown to a considerable thickness, the matrix of the rock now crystallized, forming separate crystals of microcline, plagioclase, quartz, and biotite. The new crystals of microcline, some of which are more than a centimeter in diameter, have no plagioclase mantles. This is a very significant observation; it shows, when considered in conjunction with the absence of microcline inclusions in the mantle, that the microcline crystals of the matrix were not yet in existence when the great inset completed its growth. From this it must follow that the inset grew in a liquid environment. It is difficult to believe that aqueous solutions or ions migrating through solid rock could have formed a heavy mantle of plagioclase on the large microcline and no mantles at all on smaller crystals of the same mineral, if both had been present at the same time. The microclines in the migmatite have no plagioclase mantles, indicating that they may be contemporaneous with the matrix microcline but are probably younger than the mantled inset.

The history of the big feldspar inset therefore involves early crystallization of microcline from a potassic liquid; incorporation of a few grains of early plagioclase in the growing crystal; migration of the inset into a more sodic environment where it was mantled with plagioclase; and its return to a potassic environment where the crystallization of the liquid was completed. All this is in line with the classic explanation of reversed zoning in certain crystals of effusive and hypabyssal rocks; but in this case we can go further than usual and show how the two environments came about. Walker and Mathias observed that the siltstone of the Malmesbury series contains a considerable proportion of sodic plagioclase, about An_{12}, and that the marginal granite may be as much as 10 per cent richer in plagioclase than the normal granite. The necessary sodic environment was therefore available, either generally or perhaps just locally, in the contact zone. To complete the story it is only necessary to picture the large microcline inset drifting into and again out of the contact zone. To make this possible, the essential requirement is a silicate magma dense enough to float crystals of microcline and fluid enough to permit convection currents.

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Columbia University, New York, N. Y.

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